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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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HYPOTHESES OF QUARTZ ORIENTATION IN TECTONITES

BY H. W. FAIRBAIRN

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ABSTRACT

Three hypotheses of quartz orientation in tectonites are considered, based on twinning, translation, and fracture. Emphasis is placed on the fracture hypothesis, following the ideas of Griggs and Bell. An attempt is made to apply it to tectonites showing *ac*-girdles and oblique girdles, using published diagrams and also a series of new ones. The correlation is instructive, but is only partially successful.

INTRODUCTION

The behavior of quartz in deformed rocks, or tectonites, is the subject of a large part of the literature of structural petrology. Although a considerable amount of reliable information is available, there are very few experimental data. As a result, the geologic interpretation of quartz orientation is hypothetical to a high degree. Current ideas on the devel-

opment of quartz orientation appear in publications by Sander (1930), Fairbairn (1937), Knopf and Ingerson (1938), and Griggs and Bell (1938). In the discussion which follows it is assumed that the reader has some acquaintance with this previous work.

In a recent paper (Fairbairn, 1939) the author considered the possibilities of correlation of quartz deformation with its crystal structure and was able to point out some positive relations. Twinning, translation, parting, cleavage, and fracture were all discussed. In the present paper the author considers hypotheses of orientation based on three of these types of deformation and indicates their relative importance in the light of orientations already known and of the deformation which seems to be permissible from study of the crystal structure. The opinions expressed here are not in any way final, and are intended only to promote discussion and to open the way for further investigation.

ACKNOWLEDGMENTS

During preparation of this paper the author had the advantage of association with J. F. Bell, to whom he is grateful for much advice and criticism. For financial assistance in preparation of part of the diagrams he is indebted to the Science Research Committee of Queen's University.

TWINNING HYPOTHESIS

No twinning hypothesis has ever been seriously proposed to explain quartz orientation. Of the three twins listed in Dana, two (the Dauphiné and Brazil) are much more common than the third, but unfortunately they can not be identified by ordinary optic means in a thin section. In the Dauphiné twin two right-handed or two left-handed crystals are intergrown, in the Brazil twin there is interpenetration of right- and left-handed crystals. Measurement of the vertical axis position tells nothing of the presence or absence of either of these twins. The remaining twin (so-called Japanese) is considered to be rare, but could be identified optically in an ordinary thin section. It follows a rhombohedral law in which the vertical axes of the twinned parts form an angle of 84°33′ with each other. Sander (1930, p. 43) has hinted at the possibility of twinning of this type in one example. He notes cases in which, of two adjoining recrystallized grains, the axis of one coincides with maximum II (Figure 2), whereas the axis of the other falls in the corresponding II on the other side of *ab*. The angle between the two maxima is approximately that of the Japanese twin. Since this maxima can be explained in other ways, however, evidence for twinning of this type may still be regarded as inconclusive.

Experimentally, Dauphiné twins have been produced by deformation (Zinserling, 1933), but it is not known how much volume displacement results. From the morphological character of such twinned crystals the effect would appear to be zero, since the vertical axes of the twinned

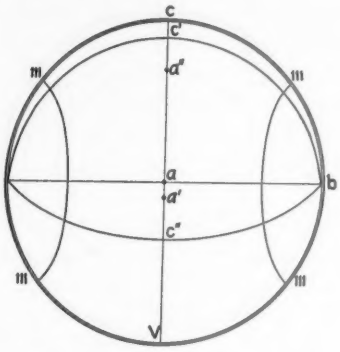

FIGURE 1.—*Hypothetical loci of quartz axes*

Assuming translation parallel to the horizontal edges [*m*:*r*]. The field of activity of this glide line is confined arbitrarily to the plane *ac*.

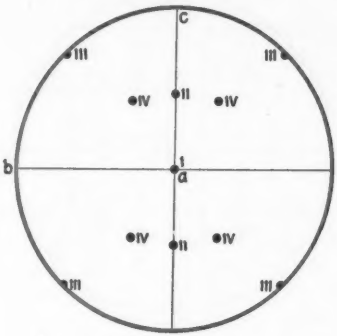

FIGURE 2.—*Quartz tectonite maxima after Sander*

Showing location of the principal quartz maxima in tectonites.

parts are coincident. However, until more is definitely known of the relation of quartz structure to twinning mechanics the importance of twins in quartz deformation will remain uncertain. In the meantime they should be looked for in tectonites and all evidence carefully reported.

TRANSLATION HYPOTHESIS

Up to the present, a majority of published diagrams of tectonite quartz have been explained by tacitly assuming a translation mechanism operating parallel to one or more directions in the grains. Sander, and especially Schmidt, have interpreted many quartz diagrams on this basis. The most common direction is believed to be parallel, or sub-parallel to the vertical axis, shown in thin section in favorable cases by undulatory extinction bands in parallel position. In other examples traces of lamellae, perpendicular to these parallel undulatory bands, are likewise believed to represent translation, but are less commonly observed.[1] Vari-

[1] Sander (1930, p. 176) has described excellent examples of quartz lamellae and has discussed their relation to mica and calcite orientation. He reasons, inductively, that the lamellae represent translation.

ous observations on "bent" quartz have also been made here and there in the literature and have been interpreted on the basis of translation. In their recent review of the literature of quartz deformation Griggs and Bell have concluded that there is as yet insufficient direct evidence in

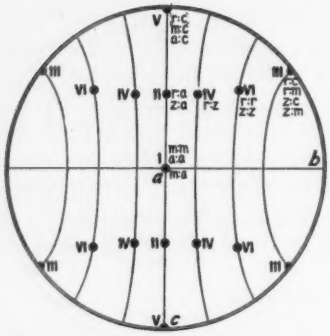

FIGURE 3.—*Quartz tectonite maxima after Griggs and Bell*

Circles joining the various maxima represent loci of crystal axes if rotation around *b* is assumed.

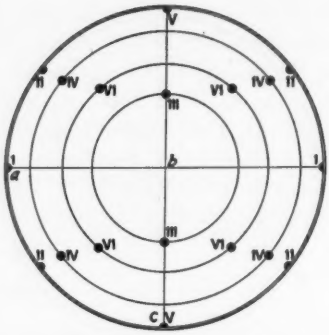

FIGURE 4.—*Quartz tectonite maxima after Griggs and Bell*

Rotated 90° around *c*.

support of translation as the principal mode of deformation. With this the author is inclined to agree, particularly in view of the crystal structural obstacles to translation-gliding (Fairbairn, 1939). The most probable glide line *t* was stated there to be [*m*:*r*] (parallel to the horizontal edges), but no specific glide plane was assigned to it. Although there is as yet no experimental evidence of translation in quartz, it is worth while to test this hypothetical glide line [*m*:*r*] for its possible occurrence in tectonites.

This is done in Figure 1, a hypothetical diagram which is to be compared with actual quartz diagrams obtained by measurement. The customary fabric axes, *a*, *b*, and *c* are used, and the glide line [*m*:*r*] is assumed to be parallel with *a*, the direction of movement in the fabric. Two cases are considered, one with a glide plane functioning, and one without a glide plane. *a* is considered as fixed, or as movable; in the latter case its freedom of movement is confined to the *ac* plane.

A. Glide line without associated glide plane.[2]

(a) With fixed *a*, crystal axis lies anywhere in *bc* plane.

[2] Although this case is not considered very probable in certain quarters, it is nevertheless included here in order to exhaust all possibilities of orientation by translation-gliding.

(b) With movable *a*, crystal axis lies anywhere in great circles intersecting in *b*. Two positions are shown in the figure having *a'* and *a''* as glide directions.

B. Glide line with associated glide plane.

(a) Glide plane is a rhombohedron lying in *ab*.

1. *a* fixed. Then crystal axes lie at maximum III.
2. *a* movable in *ac*. Crystal axes lie in small circles around *b*.

(b) Glide plane is a prism.

1. *a* fixed or movable. Crystal axis lies at *b*.

(c) Glide plane is the base.

1. *a* fixed. Then crystal axes lie at maximum V.
2. *a* movable. Crystal axes lie in *ac*.

Comparison of this hypothetical diagram with those of actual quartz tectonites (Fig. 2) shows at once that there is little similarity. Points of resemblance are the maxima found at *c*, and at III. The former is rarely found in tectonites and would correspond to maximum V (Fig. 3). The maximum at III is common but may be explained just as well by the fracture hypothesis. The *bc*-girdle of Figure 1 is duplicated in a few tectonites but may likewise be explained just as well by other means. The remaining quartz maxima and girdles are not explained at all by this assumed translation parallel to the horizontal edges. It is possible that other, as yet unknown, glide lines play a role in quartz deformation, but until there is experimental evidence of such translation, it is advisable not to be too positive in defense of this mode of deformation.

FRACTURE HYPOTHESIS[3]

GENERAL DISCUSSION

This hypothesis was first presented by Sander, and has recently been elaborated, on the basis of experimental work, by Griggs and Bell (1938). From microscopic studies of tectonites Sander believes that quartz grains are fractured and reduced in many cases to needle-like fragments whose axes are essentially parallel to the crystal axes. These needles are then rotated approximately into the shearing plane with their axes[4] parallel to *a*, the direction of movement. Measurement of the crystal axes of these needles indicates a tectonite showing maximum I. On the basis of experimental work Griggs and Bell confirmed Sander's idea of formation of needles by fracture, but found that the axes of the needles were parallel with the rhombohedral and horizontal edges as well as with

[3] Strictly, "cleavage" hypothesis is more correct. However, since Griggs and Bell have already used the general term "fracture" in its stead, and since the distinction is purely one of degree, their usage will be followed.

[4] It is important to distinguish here, and in the subsequent discussion, between needle axis and crystal axis.

the vertical edges of a crystal. The bounding faces of the needles were found to be the prism, base, and unit rhombohedron. These results were correlated with known quartz maxima by making three assumptions:

TABLE 1.—*Correlation of maxima with assumed needle axes*

Maximum	Bounding plane of needle in *ab*	Axes of needles parallel to	Number of equivalent needle orientations per grain
I	*m* or *a*	[*m* : *m*] , [*a* : *a*) , or [*m* : *a*]	1
II	*r* or *z* [5]	[*r* : *a*] or [*z* : *a*]	6
III	*r* or *z*	[*r* : *c*] , [*r* : *m*] , [*z* : *c*] or [*z* : *m*]	3
IV	*r* and *z*	[*r* : *z*]	6
V	*c*	[*r* : *c*] , [*z* : *c*] , [*m* : *c*] or [*a* : *c*]	6
VI	*r* or *z*	[*r* : *r*] or [*z* : *z*]	6

[5] In Figure 8 of the Griggs and Bell paper no distinction is made between *r* and *z*, both being indexed as (10$\bar{1}$1). As indicated elsewhere (Fairbairn, 1939), and later in this paper, the distinction has significance.

The nomenclature to describe the quartz planes referred to in the table is as follows:

c is the basal pinacoid (0001).
m is the unit prism (10$\bar{1}$0).
a is the second-order prism (11$\bar{2}$0).
r is the positive unit rhombohedron (10$\bar{1}$1).
z is the negative unit rhombohedron (01$\bar{1}$1).

The positions of these faces with respect to each other are shown in Figure 4 of the author's previous paper on quartz deformation (Fairbairn, 1939). Directions in the lattice are referred to by combining the letters for two planes whose line of intersection represents this direction—thus, horizontal edges are [*m*:*r*] or [*m*:*c*], vertical edges are [*m*:*m*] or [*m*:*a*].

(1) Needles of quartz in a tectonite are oriented with their axes parallel to *a* of the fabric. (2) One of the bounding planes of each needle lies in the *ab* plane. (3) The bounding surfaces are most commonly rhombohedrons. Figure 3 shows at I, II, III, IV, V, and VI the ideal positions of the crystal axes for quartz needles which lie parallel to *a* and which have one bounding face parallel to *ab*. The conditions for formation of each maximum are given in Table 1.

Not all of these needle orientations were found by Griggs and Bell, but the table is expanded nevertheless to include all which are compatible

with known quartz maxima and with the conclusions concerning quartz rupture given in the author's previous paper (Fairbairn, 1939).

The theoretical orientations of the axes in Figure 3 are next compared

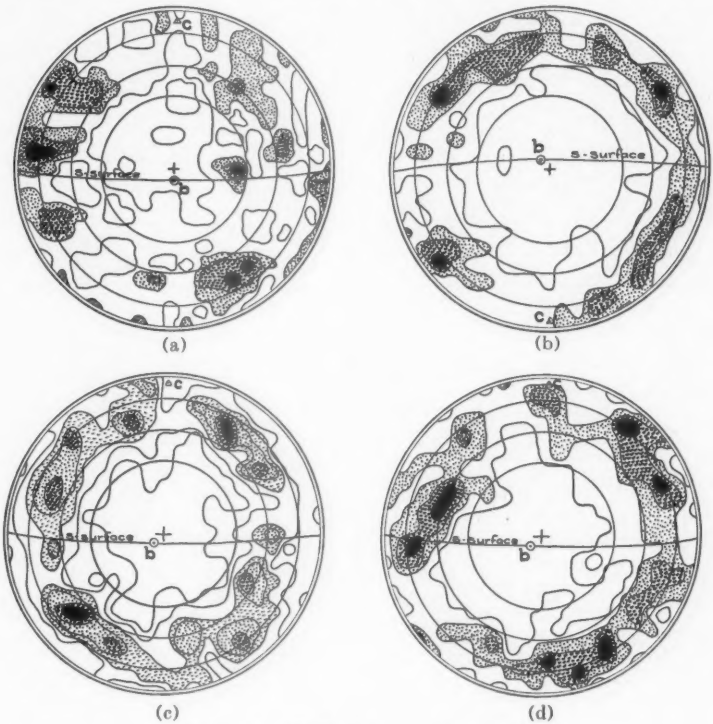

FIGURE 5.—*Girdle diagrams of quartzite*

From Brome County, Quebec. The circles correspond in diameter to those of Figure 4, and are used to test the fracture hypothesis for *ac*-girdles. Contour intervals in percent: (a) 2½-2-1½-1-½-0, (b) 4-3-2-1-0, (c) 6-5-4-3-2-1-0, (d) 4-3-2-1-0.

by Griggs and Bell with Sander's synoptic diagram of quartz maxima.[6] Sander's maxima I, II, III, and IV in Figure 2 represent the average

[6] Because of ambiguity in the use of the terms S- and B-tectonite the author prefers to drop them and to retain only the general term "tectonite". Figure 2, for example, was compiled by Sander from diagrams, some of which show single maxima (usually maximum I) and others, typical girdle concentrations. Although the former are characteristic of S-tectonites, and the latter of B-tectonites, Sander in his original synoptic diagram (1930, p. 292) states that the compilation is of S-tectonites only. It is noteworthy, too, that all published diagrams of single maxima show a tendency to formation of girdles. Furthermore, it is impossible to distinguish with certainty between S- and B-tectonites in the field. In this discussion, therefore, the distinction between the two types is unnecessary.

position of the vertical axis for a number of different tectonites. Comparison of the two figures shows the reason for retaining the same numbers. There is remarkable agreement in the locations of I, II, III, and

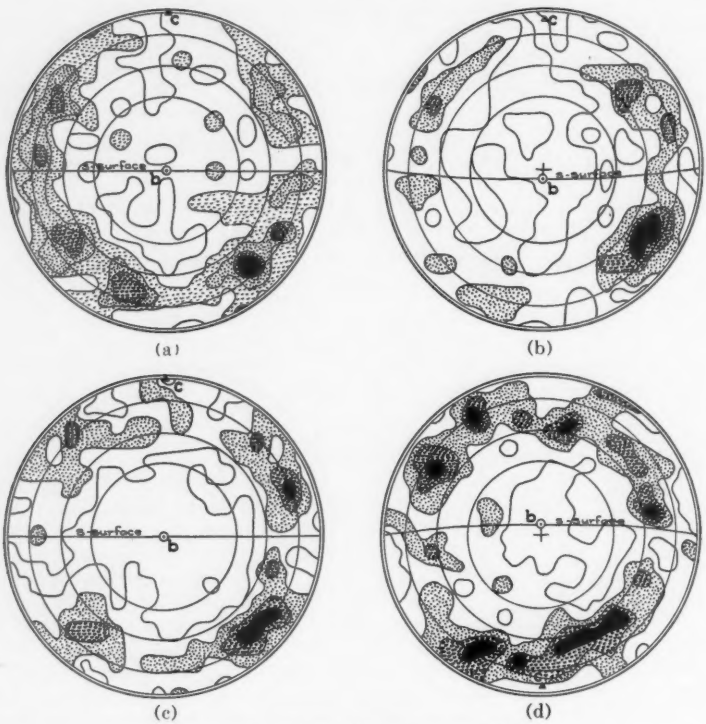

FIGURE 6.—*Girdle diagrams of quartzite*

From Brome County, Quebec. The circles correspond in diameter to those of Figure 4. and are used to test the fracture hypothesis for *ac*-girdles. Contour intervals in percent: (a) 5-4-3-2-1-0, (b) 4-3-2-1-0, (c) 4-3-2-1-0, (d) 4-3-2-1-0.

IV in both. Of the others, V occurs as a rare maximum which has been described by Sander, and VI will be discussed later.

The relation of the possible needle orientations of the table to predictable ruptures in quartz is next considered. In the preceding paper (Fairbairn, 1939) it was concluded that *r* and *z* should exhibit the best cleavage, with *m*, *c*, and *a* as second best. This conclusion was supported by experimental data. Applying these results to the Griggs-Bell hypothesis outlined above, the following predictions may be made. Max-

imum I should be common for two reasons: (1) If independent cleavages develop parallel to m and a, a considerable number of needle combinations are possible of m with m, a with a, and m with a, assuming four-

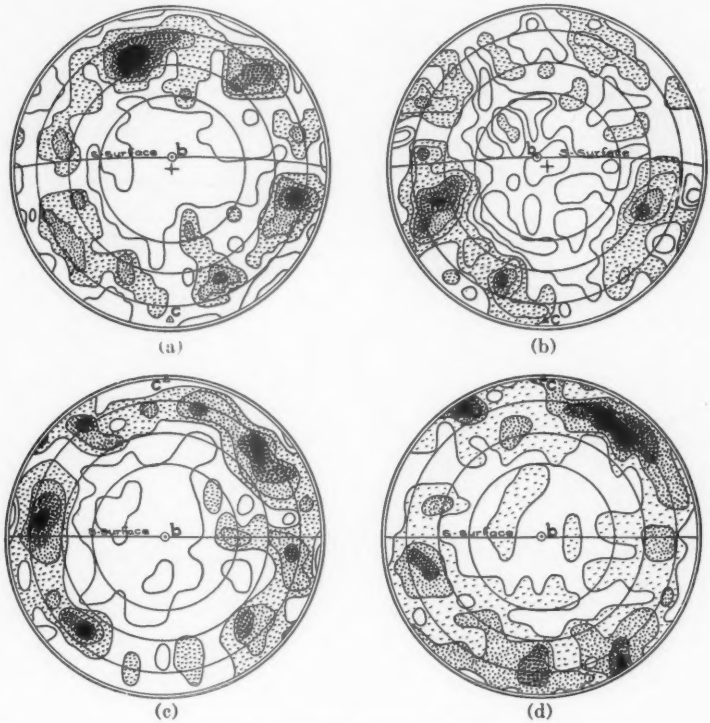

FIGURE 7.—*Girdle diagrams of quartzite*

From Brome County, Quebec. The circles correspond in diameter to those of Figure 4, and are used to test the fracture hypothesis for *ac*-girdles. Contour intervals in percent: (a) 5-4-3-2-1-0, (b) 3½-3-2½-2-1½-1-0, (c) 5-4-3-2-1-0, (d) 4-3-2-1-0.

sided needles only. If more than two sets of parallel planes co-operate to form a needle (i.e., they are polyhedral), the possibilities expand still more and a composite rupture, or linear cleavage, results. (2) If the ruptures in the crystal axis zone are composed in whole or in part of alternating r and z cleavages, a composite cleavage would result which, because of r and z participation, should be commonly developed.

Maximum III should also be of common occurrence because of the participation of r and z in all the needle combinations. Maximum II

is also dependent on r or z for one bounding face of the needles. It should not be as abundant as maximum III, because it is dependent for its development on a, a less probable cleavage plane than the m or c

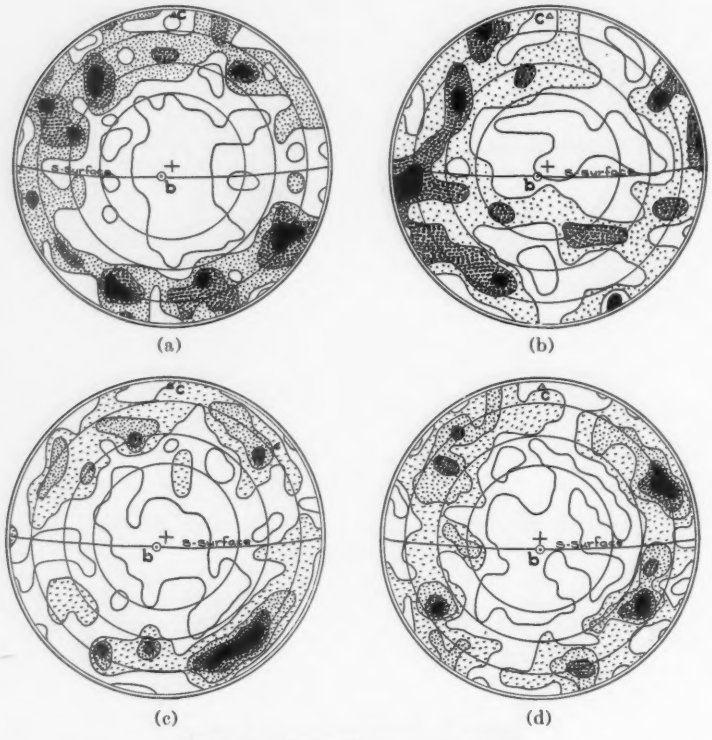

FIGURE 8.—*Girdle diagrams of quartzite*

From Brome County, Quebec. The circles correspond in diameter to those of Figure 4. and are used to test the fracture hypothesis for *ac*-girdles. Contour intervals in percent: (a) 4-3-2-1-0, (b) 4-3-2-1-0, (c) 5-4-3-2-1-0, (d) 5-4-3-2-1-0.

connected with maximum III. Similarly, maximum V should not be as common as maximum II because it is less dependent on r and z, i.e., c lies in ab and two of the four possible needle types are not bounded by rhombohedrons. For the same reason it should be less abundant than maximum III.

Maxima IV and VI are dependent on needles bounded by rhombohedrons only and would be expected therefore to be more abundant than

maxima II, III, and V. From structural considerations (Fairbairn, 1939) and from the number of equivalent needle orientations per grain which would show either of these maxima (six in each case), there is no reason

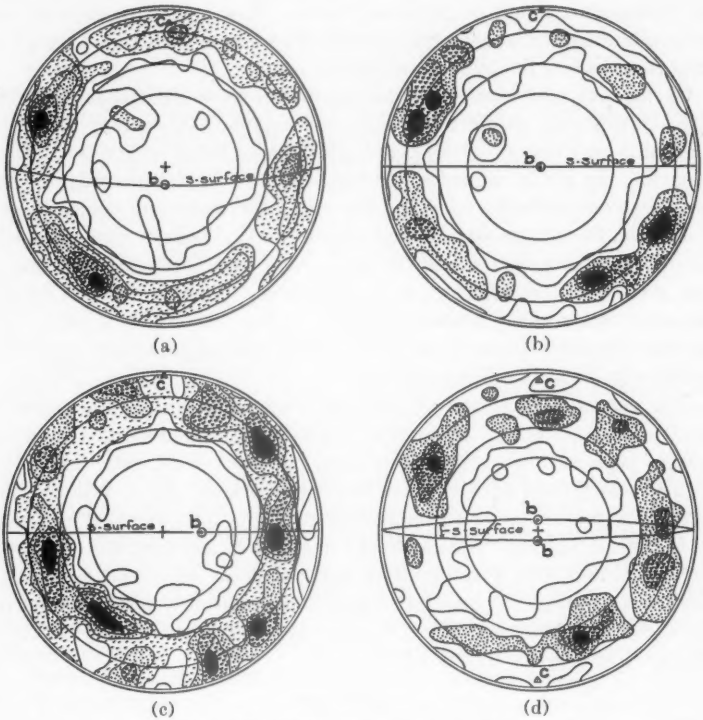

FIGURE 9.—*Girdle diagrams of quartzite*

From Brome County, Quebec. The circles correspond in diameter to those of Figure 4, and are used to test the fracture hypothesis for *ac*-girdles. Contour intervals in percent: (a) 6-5-4-3-2-1-0, (b) 4-3-2-1-0, (c) 5-4-3-2-1-0, (d) 4-3-2-1-0.

to suppose that one should occur more abundantly than the other. It is difficult to compare maxima IV and VI with maximum I, as the latter may form through development of polyhedral as well as four-sided needles. Tentatively, the following order is adopted for probable abundance of the six maxima: (1) Maxima I, IV, VI. (2) Maximum III. (3) Maximum II. (4) Maximum V.

In his synoptic diagram Sander does not include maxima V and VI. Maximum V exists (Sander, 1930, D. 36), as already stated, but is rarely

found. In the Griggs-Bell diagram (Fig. 3) maximum VI is so close to maximum IV that Sander's use of only one maximum in this region is understandable.

Griggs and Bell did not attempt a correlation of their hypothesis with the submaxima found in girdle tectonites. However, since the majority of quartz diagrams show girdles rather than single isolated maxima it is important to test the hypothesis further.

Of the different girdle orientations which have been described, two types are known to be particularly abundant. These may be called *ac*-girdles and oblique girdles. *ac*-girdles may occur singly or in pairs. If single, the girdle coincides with *ac* in Figure 3; if not single, the two parts lie symmetrically on either side of *ac*, corresponding approximately to the small circles passing through maximum IV. Oblique girdles likewise occur singly or in pairs. In the latter case they intersect in *a*, and do not show the parallelism of *ac*-girdle pairs. Their inclination to *b* is variable, but in many known examples they coincide with a great circle passing though maxima I and III. These two maxima are, moreover, commonly found in oblique girdles.

The current hypotheses of origin of these two girdle types stand in contrast to each other. *ac*-girdles of quartz axes are described from tectonites such as mica schists, phyllites, and quartzites, whereas oblique girdles are typical of tectonites such as gneisses. It is believed that *ac*-girdles develop as a result of translatory and rotary forward movement in the tectonite, perpendicular to the lineation *b*, and that oblique girdles are associated with a deformation approximately described as "flattening". In this type the translatory and rotary movements tend to neutralize each other, there is little net forward movement, and shortening occurs perpendicular to the foliation.[7]

ac-GIRDLES

The *ac*-girdles will be considered first in the attempted correlation with the Griggs-Bell hypothesis. These girdles, whether occurring singly or in pairs, are perpendicular to *b*, a fact which has given rise to the belief that rotation of various kinds around *b* plays an essential role in their development. On this assumption circles which pass through the various maxima have been constructed perpendicular to *b* in Figure 3. In Figure 4 the same construction is shown in a diagram perpendicular to *b*. It is possible in this way to test various *ac*-girdles in relation to the position of these constructed circles. This has been done for several groups of diagrams and the results are tabulated in Table 2.

[7] An alternative way of stating this is to postulate a second strain at right angles to the first, the net effect of the two giving rise to an oblique girdle.

The most conspicuous feature of Table 2 is the preponderance of submaxima which lie on the IV circle. Of the unclassified concentrations, three-quarters, at a conservative estimate, lie between the IV and VI circles. Since maxima IV and VI are the only two maxima which develop

TABLE 2.—*Statistical summary of the submaxima of ac-girdles*

Submaxima lie in circle which includes maxima	23 *ac*-girdles from Gefügekunde (Sander)	21 unpublished *ac*-girdles (Fairbairn)	20 *ac*-girdles, Figures 5 to 9	Totals	Percent
I–II–V	10	20	1	31	8
III	1	31	4	36	10
IV	56	59	52	167	44
VI	7	14	12	33	9
Unclassified	15	19	75	109	29
				376	100

through rhombohedral rupture alone, they may be added, and, if grouped with the unclassified submaxima just mentioned, give a total of 70 percent for the submaxima which lie in the region of the IV and VI circles.

It has already been predicted on grounds of crystal structure that maxima IV and VI, with maximum I, should be the most abundant maxima under the assumed fracture hypothesis. Therefore the circles in Figures 3 and 4 containing these maxima might be expected to coincide with the chief girdles of crystal axes. This is true only of the IV circle. The *ac*-girdle pairs in most of the diagrams tested coincided with the two IV circles of Figure 3. Although structurally it was predicted that maximum VI is of equal importance, relatively few submaxima lie on the VI circle.[8] Similarly, single *ac*-girdles which coincide with the I-II-V circle are uncommon, although maximum I is ranked as one of the most likely to develop. The tests show also that the III circles are relatively barren of axes. Descriptively, therefore, the *ac*-girdle diagram for a quartz tectonite is characterized by concentrations of axes centering on the circles passing through maximum IV.

Details of the third group of diagrams listed in the preceding table may be had in Figures 5 to 11. These are mostly from a folded quartzite

[8] The explanation of this discrepancy is not clear. From Table 1 it is seen that the number of equivalent needle orientations per grain is the same for maximum VI as for maximum IV. The ease of rupture is likewise the same for both *r* and *z*.

horizon in Brome County, southern Quebec, the specimens being taken at fairly regular intervals along the strike over a length of 16 miles. Several of the diagrams are from specimens of mica schist associated with the quartzite.[9] The east-dipping beds form the only foliation, and

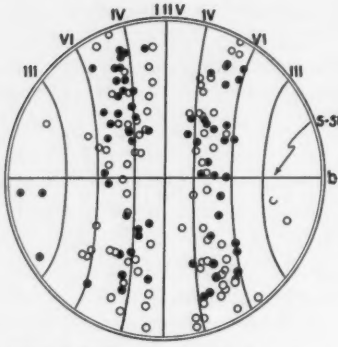

FIGURE 10.—*Summary of the maxima of Figures 5 to 9*

Rotated 90° around *c*. The filled circles represent the principal maxima; the open circles represent minor concentrations.

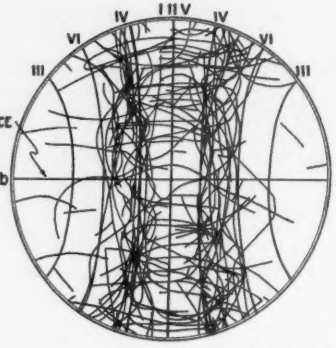

FIGURE 11.—*Trends of girdles of Figures 5 to 9*

Shown as single lines. Diagram has the same orientation as Figure 10.

there is noticeable lineation which, with few exceptions, dips steeply north. The diagrams were prepared as part of a larger correlation problem and are presented separately here because of the remarkably uniform *ac*-girdles shown over many miles along the strike. Each diagram in Figures 5 to 9 is provided with circles similar to those in Figure 4 so that the girdles and submaxima can be tested conveniently for their relation to the fracture hypothesis. Figures 10 and 11 summarize the information in two diagrams rotated 90° around *c*. Although no a priori assumption is made as to the exact location of the theoretical maxima on the various circles, it is seen from Figure 10 that the majority are clustered near and on the IV circle. Furthermore, they lie at points not far removed from the maximum IV which would be associated with the *s*-surface at the equator of the diagram. (Compare with Figure 3.)

Figure 11 shows by single lines a summary of the trends of the *ac*-girdles in the individual diagrams. The concentration in the neighborhood of the IV circle is marked.[10]

[9] This horizon is part of the Mansonville series described by T. H. Clark (1934). The quartzite occurs in thick beds enclosed in phyllite and slate, and is thought to be Upper Cambrian. The whole formation is probably not more than a few hundred feet thick.

[10] The significance of the lines crossing the *ac*-girdles at right angles is not known, but is probably not fortuitous. As the solution of this problem neither aids nor hinders the fracture hypothesis under consideration, no further reference need be made to it.

In addition to providing support for the fracture hypothesis, these quartzite diagrams demonstrate very clearly the constancy of a given type of orientation over many miles of terrane.

Three explanations have been advanced to account for *ac*-girdles, all based on *b* as a rotation axis. These are rotation of grains, rotation of strain axes, and formation of micro-flexure-folds. They are considered here very briefly.

If needles develop in *a* according to the fracture hypothesis, it is not likely that the crystal axes would retain their original orientation if the needles are subject to rotational forces. The needles would tend rather to break irregularly, or to turn so that their axes were parallel to *b*. In the latter case, the corresponding crystal axes would lie in *ac*-girdles oriented differently from any which have been described.

Rotation of strain axes about *b*, forming closed *ac*-girdles, is also improbable. It is true that a number of *s*-surfaces intersecting in *b* have been found in some tectonites, and that their development may be ascribed to change in orientation of the strain axes. Nevertheless, the angular rotation necessary to form a complete girdle is incommensurate with the probable rotations which can take place.

The third explanation, development of micro-flexure-folds, has more to commend it and, in one example, has been proven by Sander (1930, D. 162-166). An apparent obstacle to this explanation is the lack of evidence of such micro-flexure-folds in most *ac*-girdle tectonites. In the example described from southern Quebec thin sections show only the foliation marked as "*s*-surface" in figures 5 to 9. Mica flakes, where present, are not oriented so as to indicate any folding of the *s*-surfaces.

In conclusion it must be admitted that no complete explanation for *ac*-girdles can yet be offered. There is no question regarding their independent existence and their orientation with respect to the lineation. Correlation with the Griggs-Bell hypothesis also seems to be justified if *b* is regarded as a rotation axis. The girdles themselves have not yet been adequately explained however.

OBLIQUE GIRDLES

As already described, these girdles, both in their relation to *b* and in the location of the predominant submaxima, differ from the preceding *ac*-girdles. Furthermore, they have been described only from tectonites ordinarily called gneisses or granulites. Sahama's (1936) investigation of such rocks is the most complete to date. A summary of 36 of his diagrams shows that maximum I occurs 12 times, maximum II occurs twice, maximum III occurs 24 times, maximum IV occurs 10 times, maximum V occurs 3 times, and maximum VI occurs 8 times. Compared with

the table for *ac*-girdles maxima I and III are much more common here, whereas maximum IV diminishes in importance. Study of other granulite areas reveals in general the same relations.

Because no submaxima are associated with oblique girdles other than those already described, it seems reasonable to attempt a correlation with the Griggs-Bell hypothesis. The orientation of the girdles, however, precludes rotation around *b* as an explanation of the girdles themselves and of the submaxima which do not correspond exactly with the theoretical positions in Figure 3. The current explanation involves rotation about *a*, and is thought to be associated with a shortening perpendicular to the *s*-surface, in which forward movement parallel to *a* is negligible. It has been called "flattening" (Knopf and Ingerson, 1938). If it is assumed that fracturing into needles occurs first, with formation of *ac*-girdles, then it is conceivable that later slight rotation of these needles about their own axes could shift the *ac*-girdles into positions oblique with respect to *b*. Unexplained, however, is the prevalence of maximum IV in the one case and of maxima I and III in the other. Also, the scattering of the crystal axes in the oblique girdles is just as puzzling as for the *ac*-girdles and the author has no explanation at the present time.

Independent origin of oblique girdles, entirely dissociated from preexisting *ac*-girdles, is also a possibility, but encounters the same difficulty regarding their origin. In conclusion it must be admitted that girdles of all types are still to be explained, despite the impetus which the Griggs-Bell hypothesis has given to the problem.

CONCLUSIONS

The conclusions reached in the foregoing discussion may be summarized as follows:

(1) There is no direct evidence as yet in support of a twinning hypothesis of quartz orientation in tectonites.

(2) There is likewise no direct evidence in support of a translation hypothesis.

(3) All the known quartz maxima in tectonites may be correlated with the fracture hypothesis of Griggs and Bell. Both crystal structure predictions and experimental data support the correlation.

(4) The common oblique and *ac*-girdles are not adequately explained by any of the three hypotheses considered.

(5) By assuming rotation around *b*, *ac*-girdles may be partially explained by the fracture hypothesis. This extrapolation of the fracture hypothesis is successfully tested for a number of published diagrams and also for a series of new ones from Brome County, Quebec.

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MT. BAKER, A CASCADE VOLCANO

BY HOWARD A. COOMBS

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ABSTRACT

Mt. Baker in Washington is the northernmost of the major Cascade volcanoes. The actual cone measures 6000 feet in height and 8 miles in diameter. The crater rim, however, is 10,750 feet above sea level as the base of the volcano rests on top

of the Cascade range. The lavas of Mt. Baker poured out on an erosion surface of Paleozoic and Mesozoic rocks which had a relief of approximately 5000 feet.

The first flows formed a cone known as the Black Buttes, then the vent shifted 2 miles to the east and formed Mt. Baker. These lavas are noteworthy because of their lack of variation; all are pyroxene andesites. Almost all the lavas contain a type of hypersthene with inclined extinction.

The lavas from Mt. Baker and the subsidiary vents probably began to form in the Pleistocene, and activity continued until historic time.

INTRODUCTION

Surmounting the Cascade Range a series of volcanoes extends from northern California to Canada. The volcanoes in this chain, although differing slightly in form, have many characteristics in common. In petrographic and mineralogic details each is exceedingly similar. All were formed during the same general period of volcanism. Because of such similarities these volcanoes are commonly regarded as part of a system extending the length of the Cascades and possibly encircling the Pacific Ocean.

In northern California and Oregon, studies of the larger volcanoes have been made intermittently for the past 50 years, and an increased interest in volcanism during the past decade has resulted in the publication of many excellent papers.

In Washington, however, prior to 1937, the accounts of the Cascade volcanoes were limited to a few brief notes written from 30 to 60 years ago, recording the results of reconnaissance trips. It is the purpose of this paper to give more complete information on the general geology and petrography of Mt. Baker.

The closest approach to Mt. Baker is by a highway which follows the Nooksack River from Bellingham, Washington, to within 6 miles of the mountain. Another road comes within 7 miles of the volcano on the southeast side. These are the only approaches, but they are remarkable for they penetrate a comparatively wild area with a maximum relief of over 9000 feet.

The conditions for geological investigation are not ideal at Mt. Baker. Approximately half of the volcano is covered by glaciers. Most of the remaining portions are covered by dense growths of timber and brush so characteristic of western Washington. The mountains are extremely rugged, and the summer season when the snow is at a minimum is very short. The writer spent 2 months in the field during the summers of 1937 and 1938.

ACKNOWLEDGMENTS

It is a pleasure to acknowledge the many courtesies given by the rangers of Mt. Baker National Forest. The writer has profited immeasurably from discussions with his colleagues at the University of

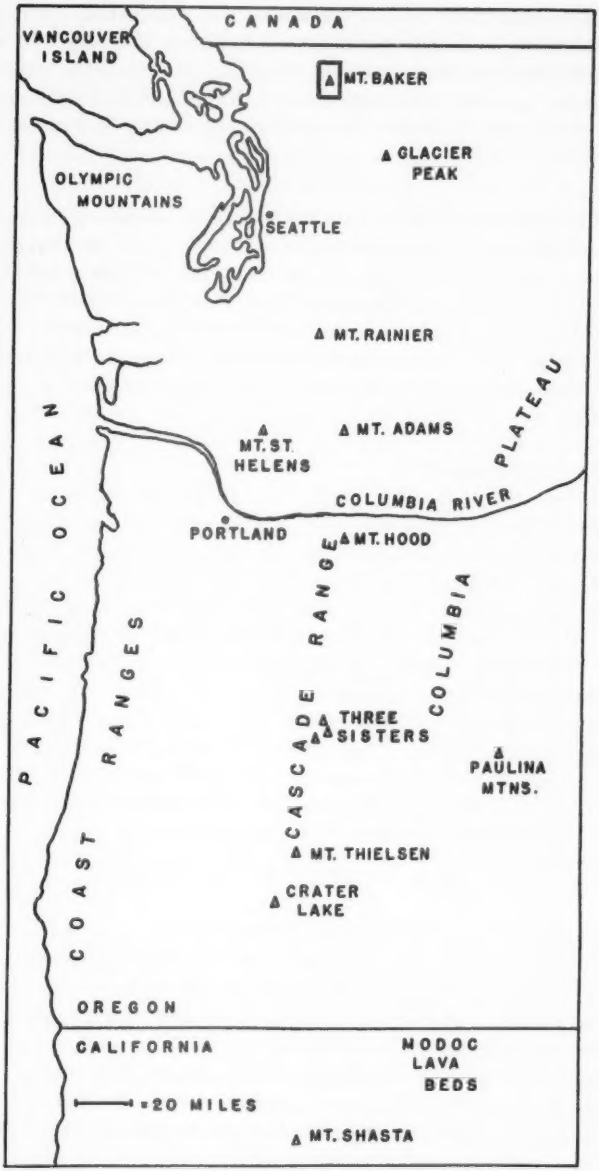

FIGURE 1.—*Sketch map showing location of Mt. Baker*

Washington and from their criticisms of the manuscript. The writer wishes to express a debt of gratitude to Dr. Howel Williams for his work on many of the Cascade volcanoes and for his criticism of the manuscript. The chemical analyses and completion of the field work have been made possible by a grant from the Penrose Bequest of the Geological Society of America.

LOCATION

The Cascade Range forms a topographic unit almost continuous from northern California to the Canadian boundary. The type of rock, however, are not coextensive. One of the most decided breaks in lithology occurs in the vicinity of Snoqualmie Pass, 75 miles south of Mt. Baker. To the south of this break the Cascades are composed of Tertiary volcanic rocks. To the north, and extending under Mt. Baker, the greater portion of the exposed rocks are pre-Tertiary. The structural trend of the rocks on both sides of the break varies from N. 70° W. to E.–W. This structural trend is almost at right angles to the north-south alignment of the entire Cascade Range.

In the immediate vicinity of Mt. Baker Paleozoic to lower Tertiary sedimentary and metamorphic rocks are the chief basement rocks. The Paleozoic portion is represented by two principal groups—greenstones and associated sedimentary rocks, and carbonaceous phyllites. Sedimentary rocks of probable Mesozoic age range from fine, thinly laminated shales to coarse conglomerates, intermediate-sized clastic predominating. The lower Tertiary is represented by a thick series of shales and arkosic sandstones containing thin coal seams. A granodiorite batholith invades some of these rocks, but its exact age is unknown.

Little is known of these formations, since all published reports dealing with them are of a reconnaissance nature. It is not in the scope of this paper to deal extensively with these older formations, but the main types of rock underlying Mt. Baker will be briefly noted.

BEDROCK BENEATH THE VOLCANO

GREENSTONE

Early reports (Smith and Calkins, 1904, p. 35) mentioned the projection of older rocks above the surface of the Mt. Baker lavas in the vicinity of Herman Mountain, due north of Mt. Baker. This older rock is part of a series containing greenstone, sandstone, and graywacke. The thickness of the series is unknown but it may be several thousand feet. The greenstones and associated sedimentary rocks vary from light gray to dark green. Sheared portions are noticeably lighter in color. Numerous veins of quartz, calcite, and epidote traverse the mass in a most irregular fashion.

Under the microscope the rock is seen to be decidedly altered. The plagioclase has been sericitized. Relics of former mafic minerals are surrounded or completely replaced by chlorite and actinolite. Diabasic textures and amygdaloidal structures suggest that the rock was of intermediate or basic igneous composition.

It is possible that the greenstone may correspond to Daly's (1912, p. 521) Chilliwack Volcanic formation of Upper Carboniferous age. The andesite in the Chilliwack Volcanic formation has been altered into typical greenstones, or, where the shearing has been particularly intense, into green schists. The type locality is only 6 miles north of Herman Mountain, and the total thickness was estimated by Daly to be 2000 feet. It is also possible that the greenstone may correspond to Daly's (1912) Vedder greenstone located 18 miles to the northwest. Recently Crickmay (1930, p. 487) assigned the Vedder greenstone to the Triassic.

CARBONACEOUS PHYLLITES

Carbonaceous phyllites outcrop at many widely scattered localities. The typical phyllites are black to slate gray. Alternating with these more carbonaceous phases are medium-gray, gritty graywackes. Less abundant are thin laminae of almost pure quartzitic sandstone. The phyllites are invariably wrinkled and often badly contorted. Slickensided surfaces are marked by innumerable quartz veins.

Phyllites have been mentioned by Bauerman (1882, p. 32; Daly, 1912; Smith and Calkins 1904) who made an exploratory trip through the district to the north of Mt. Baker. Daly's (1912) Chilliwack series (not to be confused with the Chilliwack Volcanic formation), only 5 miles north of the phyllites under Mt. Baker, is composed of slates, phyllites, and limestones, whose total thickness has been estimated to be from 10,000 to 24,000 feet. On the basis of fossils found in the limestone members of the Chilliwack series, the age has been determined as Carboniferous. It is highly probable that the phyllites surrounding Mt. Baker belong in this same general sequence.

MESOZOIC SEDIMENTARY ROCKS

It is exceedingly difficult to distinguish Paleozoic from Mesozoic sedimentary rocks unless they contain fossils, for the stratigraphy has not been studied. Fossils have been found, however, in the Canyon Creek area, 10 miles northwest of Mt. Baker, and the age of these rocks has been established as Mesozoic (Crickmay 1930)—probably upper Jurassic. Smith and Calkins (1904) provisionally assigned a Mesozoic age to some of the rocks along the Nooksack River near Mt. Shuksan because they were distinctly less metamorphosed than the Paleozoic rocks.

The writer has found probable Mesozoic rocks below the east side of Mt. Baker along the headwaters of Rainbow Creek. The rocks vary from thinly laminated, buff to dark-gray shales, graywackes, and impure sandstones to gray and white lithic tuffs and ash. Little can be said of their age, but they appear to overlie and are far less altered than the greenstones and phyllites.

TERTIARY ROCKS

A sedimentary sequence several thousand feet thick borders the western margin of Mt. Baker. The various sandstones, shales, arkoses, and thin coal seams are known collectively as the Chuckanut formation. The stratigraphy has been worked out by Weaver (1937, p. 90) who states

> "The Chuckanut formation may represent the time interval beginning in late Cretaceous and continuing into the middle or possibly late Eocene."

Daly (1992) gives the name Huntingdon formation to similar strata to the north, but this is only a provisional name to be used until a definite correlation can be made with the Chuckanut.

GRANODIORITE

Much of the area contiguous to Mt. Baker and for some distance to the northeast is underlain by granodiorite. Daly (1912) encountered this same batholith on his boundary survey and gave it the name of Chilliwack granodiorite. He thought it might have been intruded as late as the Miocene Snoqualmie granodiorite of the Mt. Rainier-Snoqualmie district. Equally probable is a Jurassic age which corresponds with numerous other batholiths in the Cascade Range. Daly has described the rock petrographically and has given chemical analyses of it.

PRE-MT. BAKER TOPOGRAPHY

Although the rocks underlying Mt. Baker are much older than those below Mt. Rainier and Mt. St. Helens, it is thought that the erosional history of these areas had much in common during the latter part of the Tertiary. The writer has dealt with the pre-Pleistocene valley cutting in the Mt. Rainier district at quite some length in a previous paper (Coombs, 1936). It will suffice to say here that even a brief visit to Mt. Baker would convince one that a rugged topography prevailed prior to and during the growth of Mt. Baker.

> Smith and Calkins (1904) report "The relations of the lava to the older terrain are interesting because they show that the lavas were poured out on a topography as rugged as at present, and that the main drainage features of the region, the Nooksack, had at the time of their eruption, the same position and even practically the same level, that it now has."

Crickmay (1930) came to the same conclusion when working in the Harrison Lake and Frazer River districts to the north. The writer has

FIGURE 1. MT. BAKER
As seen from base of Table Mountain.

FIGURE 2. TABLE MOUNTAIN FLOWS ON LEFT
Unconformably overlying the Herman Mountain greenstone, on right.

MT. BAKER AND TABLE MOUNTAIN

FIGURE 1. SUMMIT OF MT. BAKER ON RIGHT, AND BLACK BUTTES ON LEFT
As seen from the south.

FIGURE 2. PHOTOMICROGRAPH OF ANDESITE
From Rainbow Creek. Large blocky plagioclase is 4 mm. in length. Note how it is apparently composed of a grouping of smaller crystals.

FIGURE 3. PHOTOMICROGRAPH OF ANDESITE
Same as Figure 2 except under crossed nicols.

MT. BAKER AND PHOTOMICROGRAPHS

encountered no evidence about Mt. Baker to establish more definitely the time of uplift of the northern Cascades, but during the early Pleistocene the topography of these older rocks was very rugged. This conception is not compatible with many statements in the literature that the entire Cascades were peneplained during the Pliocene and that their present height is due to a major uplift at the close of the Pliocene.

MT. BAKER AND ASSOCIATED LAVAS

FORM OF THE VOLCANO

The volcanic rocks associated with Mt. Baker may be divided into three groups, from youngest to oldest: Mt. Baker, the Black Buttes, and the scattered flow remnants and subsidiary vents.

Mt. Baker is a relatively smooth, andesitic cone, little dissected by erosion. Approximately one-fourth of its 80 square miles (including the Black Buttes) is covered by glaciers. Ice has modified and accentuated the concave profile of the cone, but such dissection as exists can be measured in hundreds instead of thousands of feet as at Mt. Rainier. The base of the volcano averages 5000 feet in elevation, and the summit is 10,750 feet above sea level. Thus the actual cone is approximately a mile in height.

Most of the material that issued from the central vent was in the form of flows. These range in thickness from 15 to 50 feet. The predominating colors are medium to dark gray, black, and red. Columnar jointing is pronounced, and columns from 4 to 8 inches in diameter are most abundant. A peculiar modification of some of the columns is an offset every few inches giving the appearance of an irregular stack of books. A remarkable parallel streaking or flow structure, with alternating red and gray, or light- and dark-gray bands a fraction of an inch in thickness, is common in the Mt. Baker rocks. The thinness of the bands and their perfectly straight alignment are features worthy of more study.

Some of the flows from Mt. Baker were quite fluid especially when compared to those of most of the other large volcanoes in Washington. The longest flow, and one of the latest, follows the course of Sulphur Creek for 12 miles down to its confluence with the Baker River. An older flow reached the banks of the Nooksack River 9 miles north of Mt. Baker. This older flow may have originated in a fissure on the north side of Mt. Baker, as did the flows to the west of Austin Pass and those which poured down Bagley Creek.

Pyroclastic rocks constitute probably less than 5 per cent of the total volume of Mt. Baker, in distinct contrast to the volcanoes to the south which include abundant pyroclastic material. Tuff and pumice occur sporadically, but dark crystal ash is the most abundant pyroclastic.

At the summit are two peaks originally part of a continuous rim, now separated from each other by a crater several hundred feet deep. The north peak is a flat-topped area covered by snow and ice. Just below the ice on the west side, a narrow ledge of flat-lying red andesite can be seen in late summer. The present crater is an east-west breach in which the rocks are noticeably lighter in color than elsewhere on the mountain. The predominant colors are buff, gray, white, and red. Sulphur is extremely abundant and sublimates in cavities filling most of the porous rocks in the crater. Solfataric activity has been extremely effective in transforming much of the vent material to a mass of brilliant white opal and kaolin or in coating many of the rocks with a porcellaneous cover of opal. The south peak is composed of yellow, red, and buff tuff and flow material capped by ice and snow. The flows emerging from under the ice have an initial dip to the southward of approximately 20 degrees.

The summit of the Black Buttes (8000 feet) is overtopped by the crater of Mt. Baker approximately 2 miles to the east. The flows from the two cones meet in a saddle at an elevation of approximately 7300 feet. The Black Buttes have been greatly dissected by the Coleman and Thunder glaciers on the north and by the Deming Glacier on the south. The result is a series of sheer walls several thousand feet high, clearly exhibiting the successive flows from a central vent. These crags are so steep that little snow and ice remain on them during the summer. On account of their dark color, these crags stand out in marked contrast to the smooth ice-covered slopes of Mt. Baker.

Because the Black Buttes are so thoroughly dissected by erosion, as compared with Mt. Baker, it is assumed that the Black Buttes are the older, rather than a later parasitic cone. Flows from the Black Buttes must have been very viscous as the tongues of lava moved but a few miles from the central vent and show dip slopes as high as 30 degrees.

Scattered flow remnants occur at several places to the north of Mt. Baker. Many of these were undoubtedly derived from local vents several miles from Mt. Baker. The most accessible localities are at Table Mountain, Shuksan Arm, and Mt. Baker Lodge. These flow remnants are identical with the lavas of Mt. Baker and the Black Buttes and are at least in part contemporaneous with the larger volcanoes.

AGE OF THE VOLCANO

It is thought that the greater portion of Mt. Baker was formed contemporaneously with the other volcanoes to the south during the Pleistocene. The latest eruptions from Mt. Baker have been recorded within historic time. Gibbs (1870, p. 140) stated,

> "I am informed both on the authority of the Hudsons Bay Co., and also the Indians, that the eruption of 1843 was the first known. It broke out simultaneously with Mt. St. Helens and covered the whole country with ashes."

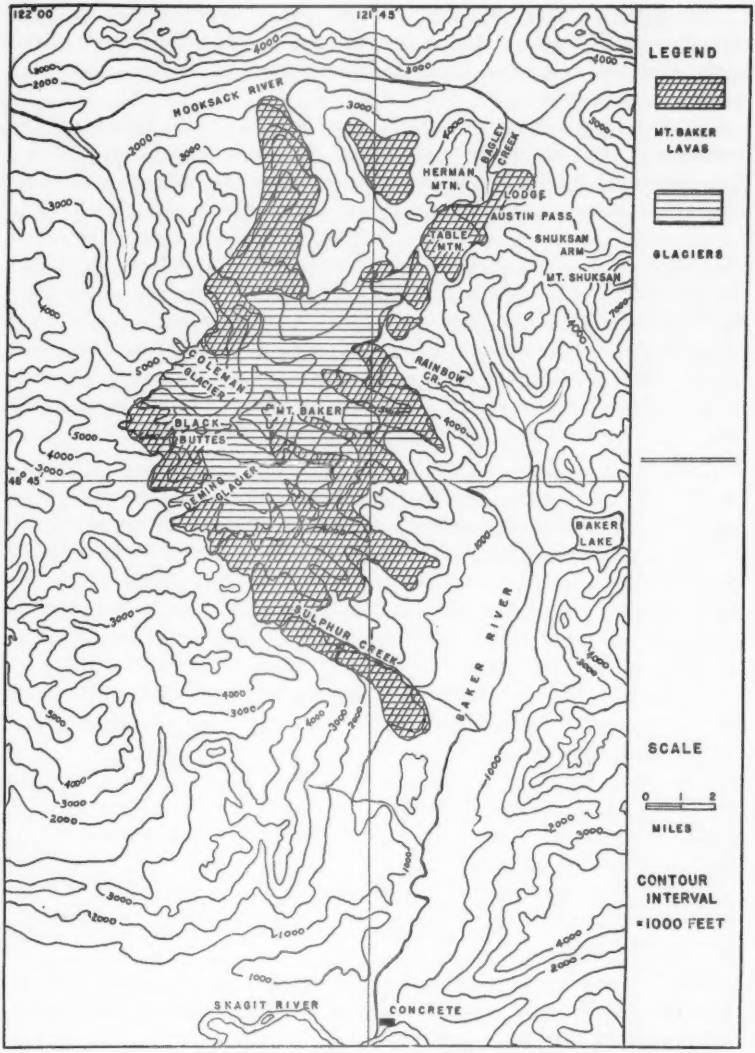

FIGURE 2.—*Geologic sketch map showing distribution of Mt. Baker andesites*

Diller (1915, p. 56) mentions that eruptions also occurred in 1854, 1858, and 1870, but none of these eruptions included lava flows; ash alone was ejected.

The Black Buttes probably had a much earlier beginning than Mt. Baker. The Black Buttes also seem to be older than most of the other Cascade volcanoes (Mt. Thielsen excepted) if the amount of erosion is a criterion.

The small remnants of flows to the north of Mt. Baker present evidence of both early and late origin. On Shuksan Arm, overlooking Mt. Baker Lodge, andesitic remnants cling to the side of a severely glaciated valley several hundred feet above the valley floor. This would represent one of the earliest flows, either of preglacial or interglacial age. In contrast, the bottom of the valley in which Mt. Baker Lodge is situated is covered by an extremely fresh flow of andesite which definitely did not pour out until after the valley was glaciated. It is impossible to correlate these flow remnants with either Mt. Baker or the Black Buttes, but undoubtedly they are all phases of the same general period of volcanism.

PETROGRAPHY

General considerations.—Petrographically the lavas from Mt. Baker are so similar to those of Mt. Rainier that little need be given here regarding their mineral content and texture. A more complete account of the petrographic details has been published (Coombs, 1936). In this discussion only the salient points of the Mt. Baker lavas will be discussed, and an attempt will be made to correlate, as much as possible, the petrographic character of the rocks with their relative positions in the volcano.

Hague and Iddings (1883) were the first to appreciate the variety of lavas encountered in the southern Cascade volcanoes as compared to the lack of variation in the lavas in the northern Cascades. Although they were not acquainted with Mt. Baker, this volcano fits well into their general impression of the Cascade volcanoes. Mt. Baker is composed exclusively of andesite, and of this amount 90 per cent is pyroxene andesite.

Rather than describe a few specimens in detail, the writer will describe each mineral in order of its importance, and mention any unusual features.

Plagioclase.—Smith and Calkins (1904, p. 58) described a specimen of andesite from Austin Pass and were impressed by the size relationships of the plagioclase crystals. They state

> "At the first preliminary examination of the thin section through the low power lenses, the writer was struck by the fact that the feldspars were divisible on the basis of size into three classes. Speaking generally, the first class consists of those crystals easily recognized in the hand specimen by the unaided eye; the second, of those that produce the reflections that give the groundmass its sparkling crystalline appearance; the third, of those of purely microscopic size."

These three grades of feldspar crystals are common to practically all of the Mt. Baker lavas, although the quantity of phenocrysts, as compared to the groundmass, ranges between wide extremes. The largest phenocrysts are thought to be due to a cumulophyric clustering of smaller ones. Evidence for this is the glassy inclusions which mark out possible individual crystals, comparable in size with the intermediate-sized plagioclase crystals, and the extremely complex twinning, probably resulting from the heterogeneous grouping of partially resorbed phenocrysts.

The largest crystals average 3 mm. in length, are complexly twinned, strongly zoned, and often full of brown, glassy inclusions. The usual types of zonal structures show calcic cores surrounded by more sodic rims. The progression from core to rim is gradual, but often there are interruptions and oscillations. This might be considered "normal" according to Homma (1936, p. 140) and Larsen (1938, p. 229). The amount of variation in a single crystal is usually between Ab_{60} and Ab_{54}.

The intermediate-sized plagioclases average .5mm. in length. They are characteristically tabular and commonly show Carlsbad twinning; polysynthetic twinning is not always distinguishable. The composition ranges from Ab_{70} to Ab_{41}.

The smallest crystals range from microlites to .07mm. They are typically elongated, often with castellated terminations. In composition, lack of inclusions, and twinning the smallest plagioclase crystals are identical with those of intermediate size.

Hypersthene.—Almost all of the Mt. Baker rocks contain hypersthene. The most unusual feature of this mineral is its "monoclinic" tendency. At least one crystal of hypersthene with a definitely inclined extinction was encountered in every section of Mt. Baker lavas examined. The angles range from 1 to 15 degrees, although by far the greatest number have maximum angles of 8 degrees. Occasionally zoned crystals were observed in which the extinction angles varied from center to periphery by as much as 5 degrees.

For a more complete description of the "monoclinic" hypersthenes which occur in other Cascade volcanoes, the reader is referred to Coombs (1936), Verhoogen (1937), and Thayer (1937). This type of hypersthene has little in common with the varieties described by Winchell (1933) and Hess and Phillips (1938, p. 453).

The pleochroism in both ortho and clino varieties is distinct: X = orange, Y = yellowish brown, and Z = green. The optic angle changes considerably depending in part upon the amount of magnetite included

in the crystals. In those heavily charged with magnetite, the 2V drops as low as 60°, but in the more common clearer crystals, 2V averages 70°.

Occasionally augite has crystallized in jackets around the partially resorbed hypersthene so that both pyroxenes have their *c* axes and their prismatic cleavages parallel. A similar structure was observed by Williams (1933) at Mt. Thielsen, by Coombs (1936) at Mt. Rainier, and by Thayer (1937, p. 1624) in the lavas of the Sardine series in the Oregon Cascades.

Augite and pigeonite.—The pyroxenes are far less abundant than hypersthene. Pigeonite with a $2V = 40°$ was seen in only five specimens. Augite is common to almost all the lavas. After the abundance of augite in 100 sections picked at random was plotted it was evident that augite is far more plentiful and forms definitely larger phenocrysts in the earlier, more fluid flows. The average length of the augite crystals in all the lavas is .5 mm. The diopsidic nature of the augite is indicated by the following optical properties: $\alpha = 1.690$, $\beta = 1.700$ and $\gamma = 1.720$; the optic angle varies little from 60°; $Z \wedge c = 40°\text{-}60°$.

The peculiar association of pigeonite and olivine (now partially altered to antigorite) and hypersthene was noted in a specimen from the south side of Table Mountain. The only other reference to such an association of minerals is by Verhoogen (1937, p. 65).

Olivine and hornblende.—One of the most striking characteristics of the Mt. Baker lavas is the presence of olivine in most of the early flows. It is not to be inferred, however, that olivine is abundant in the Mt. Baker rocks for seldom does it make over 10 per cent of the rock. Its occurrence in the long tongues of early valley-filling flows is rather to be expected as these flows were far more fluid and probably hotter than the later viscous flows of Mt. Baker. Most often associated with the olivine are hypersthene and augite. A few slides show both hornblende and olivine as well as the pyroxenes, but in these the hornblende is resorbed or almost entirely replaced by magnetite. The perfectly euhedral outlines of the replaced hornblende leave little doubt as to its identity.

Both basaltic hornblende and ordinary hornblende are present. The ordinary type is more abundant and shows the following pleochroism: X = golden yellow, Y = olive brown, and Z = deep golden brown; $Z \wedge c = 20°$. The hornblende always presents euhedral outlines. Occasional hornblende crystals were perfectly fresh, others were rimmed by magnetite, and some were completely replaced by magnetite.

Types of groundmass.—The general appearance of the Mt. Baker lavas is dependent mainly upon the groundmass. The more glassy varieties have a distinctive brilliance in color and lustre; they are usually a shiny

black, dark brown, or hematite red. In contrast, the less glassy rocks are usually intermediate shades of gray to light gray. All are charged with liberal quantities of magnetite or hematite dust, sometimes in sufficient quantities to render the thin sections opaque.

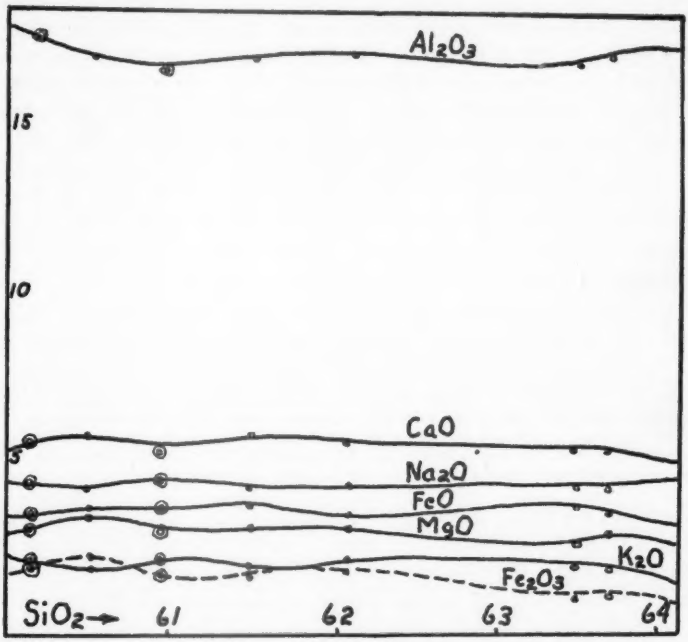

FIGURE 3.—*Variation diagram of the Mt. Baker and Mt. Rainier andesites*
(The analyses with double circles are from Mt. Baker)

Approximately 90 per cent of the Mt. Baker rocks have a hypocrystalline groundmass. The amount of glass, however, varies within wide extremes. The remaining 10 per cent is mostly in the holohyaline group. This is especially true of the dark pumice and ash. Holocrystalline rocks are rare.

CHEMICAL COMPOSITION AND INTERPRETATIONS

Variation diagram.—As only two analyses are available from Mt. Baker, little can be said about the chemistry of the volcano. However, several analyses heretofore unpublished are now available from Mt. Rainier, and these will be included in the accompanying variation diagram.

The lack of variation is the most striking point in connection with these analyses. All specimens analysed were collected from widely separated localities and represent lavas extruded at different times. At Mt. Rainier it is probable that the lavas were extravasated in rather rapid

TABLE 1.—*Chemical analyses of andesite*

	1	2	3	4	5	6	7
SiO	60.16	60.95	60.53	63.50	61.50	63.57	62.08
Al_2O_3	17.65	16.87	17.01	16.91	17.14	17.06	17.25
Fe_2O_3	1.92	1.58	2.12	1.12	1.55	1.17	1.80
FeO	3.44	3.52	3.58	3.61	3.85	3.32	3.32
MgO	2.90	2.75	3.47	2.28	2.96	2.78	2.91
CaO	5.51	5.20	5.79	5.14	5.80	5.11	5.43
Na_2O	4.25	4.48	4.01	4.14	4.11	4.21	4.18
K_2O	1.84	2.08	1.58	1.84	1.65	1.72	1.84
H_2O	.83	.97	.25	.09	.20	.09	.16
H_2O	.19	.16	.12	.05	.04	.09	.09
TiO_2	.76	.90	.95	.81	.91	.18	.20
P_2O_5	.26	.27	.26	.24	.25	.73	.85
MnO_2	.11	.10	.10	.09	.10	.08	.08
	99.82	99.83	99.77	99.82	100.06	100.11	100.19

1. Andesite from the north side of Mt. Baker at an elevation of 8300 feet. R. B. Ellestad, analyst.
2. Andesite from Austin Lake, Mt. Baker, R. B. Ellestad, analyst.
3. Andesite from Burroughs Mountain, Mt. Rainier. A very late flow. R. B. Ellestad, analyst.
4. Andesite from the fissure at St. Elmos Pass, Mt. Rainier. R. B. Ellestad, analyst.
5. Andesite from the summit of Mt. Rainier, at Register Rock. R. B. Ellestad, analyst.
6. Andesite from Paradise, Mt. Rainier. A flow intermediate in age. R. B. Ellestad, analyst.
7. Andesite from Mt. Rainier, a locality near the terminus of the Nisqually Glacier. One of the early flows resting unconformably on the granodiorite. R. B. Ellestad, analyst.

succession, allowing little time for differentiation. At Mt. Baker topographic evidence indicates that more time elapsed between the earliest and latest flows. Both analyses from Mt. Baker are from relatively late flows. Petrographically a small difference does exist between the earliest and latest flows as there is a definite decrease in the amount of olivine in the younger lavas.

The lavas from Mt. Baker and Mt. Rainier are slightly different from the andesites of the volcanoes to the south. The Mt. Shasta andesites, for example, are richer in alumina and magnesia and lower in potash and ferrous iron than the lavas from Mt. Baker and Mt. Rainier. At Lassen Volcanic National Park the andesites average 2 per cent lower in silica and approximately 1.5 per cent lower in potash and soda. Ferric iron, magnesia, and lime are over 1 per cent higher. The Mt. St. Helens andesites are approximately the same as the Mt. Baker andesites. It is noteworthy that the volcanoes to the south range in composition from rhyolite to basalt, and the intermediate types are especially abundant. In the volcanoes to the north the lavas are restricted to variations of andesite.

Interpretations.—The problem of volcanism in the Cascade Range is a broad one, and most writers believe that it is intimately associated with

the Cenozoic volcanism which characterizes the major portions of Washington, Oregon, and northern California.

Several writers have discussed the relationships of the lavas in the High Cascades, as defined by Callaghan (1933, p. 243), with those of the vast Columbia Plateau to the east. Williams (1935, p. 299) suggested two possibilities for the high calcic nature of the High Cascade lavas as contrasted to the calc-alkaline Columbia Plateau flows: (1) The High Cascade volcanoes tap lower levels in the primary basalt, and in these lower levels there is a concentration of early crystals; (2) the magmas beneath the High Cascades may have assimilated large amounts of shale, limestone, and dolomite, thus accounting for their rich lime content. In connection with Williams' first point, there is little direct evidence. It is unknown whether the source magma for these volcanoes was a differentiate caused by crystal settling of early, basic plagioclase feldspars and magnesian olivines and pyroxenes or whether it was due to some other cause. It is quite certain, however, that if differentiation was responsible for this source magma, this process was brought to a sudden halt when the lavas of Mt. Baker and Mt. Rainier poured forth on the surface. The conditions during the time of extrusion of these lavas must have been very stable, or the time between the earliest and latest flows was exceedingly short, for the lavas of these two volcanoes show practically no evidence of differentiation. In regard to the second point, the rocks beneath these volcanoes contain abundant lime and alumina in the form of calcareous shales, phyllites, and calcareous arkose. These underlying rocks may have had a decided influence on the ultimate character of the lavas. If these lavas represent contaminated fractions or differentiates of a regional primary magma (similar to the Columbia Plateau basalts) the volumetric proportions are also important. The andesites of both Mt. Baker and Mt. Rainier total less than 100 cubic miles. Contrasted to this, the Columbia basalts total over 100,000 cubic miles.

Van Bemmelen (1938) studied an area in the Soenda Mountains of Java which is almost identical to the Mt. Rainier district. His theory as to the origin of these magmas involves the magmatization of preexisting crystalline rocks by ascending emanations from the substratum (secondary or palingenetic magmas) and hybridization or contamination of juvenile of palingenetic magmas by assimilation of foreign material (syntectic or mixed magma). This theory might also apply to the northern Cascade volcanoes.

Callaghan (1933) and Thayer (1937) have pointed out many differences between the lavas of the High Cascades and those of the Western Cascades in Oregon. Wells and Waters (1935, p. 971) have demonstrated a marked difference between the Eocene lavas and those of the Western

Cascades in Oregon. More recently Waters has made a study of the Cascade Mountains and the Columbia Plateau and has drawn some striking relationships between the various magmatic groups within this general province.

It is the opinion of the writer that Mt. Baker is but a unit in a chain which may include all the major Cascade volcanoes. There seems to be a very definite relationship between the structural conditions prevailing in each unit of this chain and the chemical and petrographic character of the associated volcanic rocks.

It is possible that the lavas and pyroclastic rocks of Mt. Baker may represent differentiates of a regional primary magma. Evidence for or against this theory can be obtained only by careful analysis of the petrographical, chemical, and structural data available. The evaluations of interpretations would remain a difficult problem even if these data were more abundant.

Evidence does indicate, however, that both Mt. Baker and Mt. Rainier were formed from magmas almost identical in composition, and during the growth of the cones there was little differentiation.

The condition and kinds of rock underlying these volcanoes may have a decided influence on the chemical composition of the lavas in these cones. The necessary petrologic and structural data on the underlying rocks are quite meager, and few chemical analyses are available.

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DEPTH AND GEOGRAPHICAL DISTRIBUTION OF DEEP-FOCUS EARTHQUAKES

(SECOND PAPER)

BY B. GUTENBERG AND C. F. RICHTER

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INTRODUCTION

This paper revises and extends the results reported by Gutenberg and Richter in 1938. New data used include: (1) Those of the International Seismological Summary from April, 1932 to June, 1933; (2) a valuable catalogue by S. P. Lee and others, based principally on the records at Chiufeng, China (1938); (3) bulletins of the Jesuit Seismological Association; (4) bulletins from the international seismological office at Strasbourg; (5) bulletins of groups of stations, particularly those making determinations of epicenter and depth, such as the United States Coast and Geodetic Survey, the Russian group, the Earthquake Research Institute (Tokyo), the New Zealand stations, the Indian stations, and Batavia, Dutch East Indies, with its auxiliary stations; (6) bulletins of individual stations, especially those equipped with vertical-component instruments which report their readings in detail; (7) original seismograms written at Huancayo, Peru (up to August 1937), Pasadena, California and auxiliary stations, and Tucson, Arizona (since April 1937, by courtesy of the United States Coast and Geodetic Survey); (8) a new list by Berlage (1939). In many instances determination of depth has depended exclusively on the original seismograms available. Otherwise,

the general procedure has been that described in the previous paper. Very often the precision of the entire solution depends on the correct identification of *pP*.

REVISION OF PREVIOUS RESULTS

The receipt of new data, and the discovery of errors, have led to the following revisions to be applied to the determinations published in 1938. Numbers are those of the catalogue of shocks there included.

No. 52. A detailed paper on this shock has been published by Brunner (1938). He gives 25° S. 179.4° E., depth 600± km. The shock is multiple. The origin time (16:09:40) given in the catalogue of Gutenberg and Richter refers to the first shock; Brunner has worked chiefly with the very large later shock, for which he gives 16:09:58±. The International Summary gives 24.0° S. 179.2° E., depth 0.05 R, origin times at 16:09:19 and 16:09:29. Re-examination of all data leads to no significant modification of previous determination.

No. 107. Further investigation suggests a depth of about 700 km (instead of 650).

No. 115. This shock has been newly discussed in detail by H. P. Berlage, Jr. (1939). He gives some additional data and finds 4.7° South, 111.3° East, $0 = 23{:}03{:}40$, $h = 600$ km. A revision using all available data leads us to 4° South 111° East, $0 = 23{:}03{:}42$, $h = 620$ km. The difference in the results is partly due to use of different travel time curves.

For valuable discussion and additional information on the following three shocks in the Philippine Islands the authors are much indebted to Father Repetti (personal communication).

No. 129. Since macroseismic reports of this shock came only from eastern Samar and eastern Leyte, in the Philippines, Father Repetti put this epicenter at about 12° N. 126° E. However, this would lead to large residuals with a clear distribution in azimuth. As the distribution of macroseismic effects is often very irregular in deep-focus shocks (this is especially well established in Japan), the authors prefer to make no change.

No. 133. All macroseismic reports are from Samar, Leyte, and northern Mindanao. Father Repetti's epicenter is 10½° N. 126° E. This would lead to large residuals at a few stations. An epicenter between this and the position adopted by the authors (9½° N. 123° E.) would fit most of the observations very well. Accordingly, the authors offer the following revised epicenter: 9½° N. 124¼° E. Origin time and depth are unchanged.

No. 135. Father Repetti prefers 15° 10′ N., 119° 40′ E., which is about one degree distant from the epicenter used by the authors.

No. 180. Origin time: For 09:41:21 read 09:41:51. No. 184 is another solution for the same shock, duplicated accidentally.

No. 183. This determination has been revised after receiving data for the Japanese stations, through the International Summary. The revised solution is as follows: April 4, 1932, 19:16:38, 30° N. 139° E., 410 km. AAAakp. This shock has been investigated by Jeffreys (1939), who finds 30.38° N., 138.79° E., depth 0.06132 R ± 0.00053 R.

No. 184 should be canceled. (*See* No. 180 above.)

No. 211. This has been revised using the data of Japanese stations from the International Summary, with the following result: Sept. 23, 1932, 14:22:11, 44¾° N. 138° E., 300 km. AAAak.

Other shocks have been examined using additional data, but without leading to significant changes.

In the previous paper (1938) the authors referred to a partly empirical formula by Gassmann for evaluating the focal depth when the distribution of macroseismic intensity is known; they also used a modification of this formula suggested by Blake. In a recent note, Gassmann (1938) states that Blake's formula is the correct one, and that the discrepancy is due to a misprint in Gassmann's original publication (1927). Consequently, in the previous paper equation (2), and the column headed h' in Table 4, should be canceled.

There are three additional recent shocks for which macroseismic data are available and to which this method can be applied. Data and results are as follows:

No.	Date	Region	R(km.)	J	h''(km.)	h(km.)
5n	Feb. 5, 1938	Colombia	1200±	7	150	160
250d	Nov. 14, 1937	Hindu Kush (Central Asia)	1300	8	100	240
252p	July 13, 1938	Rumania	350±	4½	110	150

Numbers are the additional catalogue numbers of Table 1 of the present paper. R is the radius of the shaken area, J is the maximum intensity, h'' is the depth calculated from Blake's formula, and h is the depth from macroseismic data. The discrepancy for the Hindu Kush shock may be due either to incomplete data or to the imperfection of the method at larger depths.

CATALOGUE OF DEEP-FOCUS EARTHQUAKES

The new determinations are given in Tables 1 and 2. The arrangement is that of the first paper (1938); the geographical order has been preserved by assigning to each of the new shocks a number and letter placing it between the shocks of the original list. Shocks in Mexico north of the previous No. 1 have been given a number 0 and corresponding letter.

TABLE 1.—*Data for deep-focus earthquakes*

[First three capital letters in last column refer, in order, to quality of determination of epicenter, origin time, and depth—(A) very accurate; (B) good; (C) fair; (D) poor. Small letters refer to comparison with other sources, especially the International Seismological Summary (I. S. S.)—(a) at least approximate agreement with the Summary as to epicenter and depth; (b) I. S. S. gives shock as normal, with epicenter within 4°; (c) I. S. S. gives shock as normal, epicenter more than 4° distant; (h) I. S. S. not available; (n) Berlage gives approximately same epicenter and depth; (p) used in "Materials" I or II (Gutenberg and Richter, 1936, 1937); (g) Lee gives approximately same epicenter and depth.]

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
Mexico, Central America, Caribbean Region.						
0 d	1933, Jan. 24	15:39:09	18¾N	101¾W	90	AABb
0 g	1938, June 28	19:17:42	18 N	100 W	110	ABAh
0 j	1938, May 3	02:15:27	18½N	99 W	100	AAAh
0 k	1937, Oct. 6	09:47:18	18½N	99 W	100	BAAh
0 n	1937, July 26	03:47:11	18.4N	95.8W	100	AAAh
0 q	1937, May 28	15:35:53	17 N	93 W	150	CBBh
0 t	1937, Jan. 11	13:21:16	16 N	94½W	110	BBAh
1 d	1935, Mar. 17	21:33:18	14½N	92 W	110	BAAh
1 g	1937, Dec. 5	05:42:09	14 N	91 W	80	CCAh
1 i	1939, Jan. 20	20:40:27	13½N	91½W	70	BBBh
1 n	1932, June 20	09:01:47	12½N	89 W	80	AAAa
1 r	1932, May 21	10:10:07	12 N	87½W	90	AAAa
South America, intermediate shocks.						
5 k	1935, Sept. 18	04:58:00	5½N	76 W	80	BBBh
5 n	1938, Feb. 5	02:23:34	4½N	76¼W	160	AAAh
5 q	1935, Oct. 27	22:05:05	4 N	76 W	150	BCBh
6 m	1937, May 21	13:12:25	2½N	77½W	90	AAAh
7 m	1937, July 19	19:35:24	1½S	76½W	190	AAAh
8 c	1935, Nov. 2	21:05:40	2 S	79 W	130	CCCh
8 m	1935, Mar. 8	11:59:24	4 S	80 W	100	CCCh
8 r	1936, May 6	03:38:55	8 S	75 W	160	BBAh
9 m	1937, June 22	05:34:10	9 S	80 W	150	DDDh
10 d	1935, Mar. 9	02:54:37	12 S	78 W	150	CCCh [1]
10 p	1937, Mar. 28	12:21:42	13½S	75 W	120	CCCh [2]
11 d	1935, Sept. 19	09:55:47	15½S	70 W	250	BCBh
11 m	1937, Mar. 29	07:49:47	15½S	71 W	120	BBAh
11 r	1935, Mar. 26	19:54:47	15½S	73 W	120	BBBh
12 k	1937, July 9	17:27:40	16 S	72 W	180	BCCh
12 m	1936, Sept. 16	17:44:13	16 S	72½W	130	BBAh
12 r	1936, July 4	08:52:35	18 S	70 W	140	BBAh
13 k	1932, July 29	00:45:18	19 S	70 W	110	BAAb
13 p	1934, Dec. 4	17:24:38	19½S	69½W	130	BBCh
13 s	1936, Dec. 5	00:37:58	20 S	70½W	100	BBAh [3]
15 a	1934, Dec. 23	09:52:28	21 S	68 W	100	BBAh
16 d	1936, Jan. 31	15:14:15	22 S	67 W	160	CCCh
16 m	1936, Nov. 29	14:55:10	22½S	67 W	230	BCCh
16 r	1936, June 22	10:28:06	22 S	68 W	100	CBAh
16 u	1931, July 18	05:27:05	22½S	69 W	150	BABb
17 k	1937, Feb. 24	18:45:45	23 S	67 W	260	CCCb [4]

[1] Huancayo has *P* at 02:55:23, *S* at 02:55:51.
[2] Huancayo has *iP* at 12:22:16, *iS* at 12:22:42.
[3] Aftershock at 04:19:42.
[4] Huancayo has *P* at 18:48:57, *S* at 18:51:24.

TABLE 1.—*Data for deep-focus earthquakes—Continued*

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
South America, intermediate shocks—*Continued.*						
17 l	1936, Nov. 7	06:32:47	23 S	67 W	200	BBBh [5]
17 m	1935, Feb. 28	07:10:25	23 S	67 W	200	CCCh
18 m	1933, Feb. 10	08:46:01	23 S	68 W	110	CCBa
19 m	1938, Aug. 4	08:54:51	23¼S	66¾W	220	BAAh
20 m	1936, Nov. 7	05:07:30	24 S	66 W	200	CCCh [6]
21 i	1935, Sept. 28	04:00:27	24 S	67 W	150	CCCh
21 p	1934, Dec. 16	16:31:10	24 S	68 W	150	CCCh
22 m	1932, Nov. 1	10:38.9	24 S	70 W	100	CCBc [7]
23 d	1937, Mar. 14	11:55:48	24½S	69½W	60	BBAh
23 i	1935, Feb. 13	17:22:01	25½S	69 W	100	BAAh
23 m	1936, Dec. 19	02:57:37	28½S	68½W	160	CCBh
23 p	1937, Mar. 19	18:11:55	29 S	70 W	70	BCBh
23 r	1936, April 30	17:06:17	31 S	67 W	200	BBBh
23 w	1938, June 15	07:43:50	31 S	70½W	70	BBAh
24 d	1932, Nov. 29	11:11:05	32 S	71 W	110	BBAa
24 g	1932, May 8	19:21:27	32 S	70 W	110	DDDb
24 h	1932, May 10	11:53:04	32 S	70 W	110	DDDb
24 m	1937, Feb. 12	18:32:58	32 S	66½W	200	BBBh [8]
24 r	1937, Jan. 30	10:31:52	36 S	72 W	100	CCBh
24 t	1939, Jan. 25	03:32:14	36¼S	72¼W	70	AAAh
South America, very deep shocks.						
31 a	1935, Dec. 28	04:50:49	9½S	70½W	650	AAAh [9]
39 a	1934, Jan. 9	07:32:32	28½S	63 W	630	BBBh [10]
New Zealand.						
41 f	1938, Oct. 30	12:45:23	38½S	176½E	150	BCCh [11]
Kermadec-Samoa region, intermediate shocks.						
41 m	1937, Sept. 1	08:38:59	32 S	180	120	BBCh
41 s	1932, July 20	20:05:46	27 S	178 W	170	BAAb
43 m	1934, Nov. 9	03:58:56	17½S	174 W	80	BBBh
44 m	1935, Aug. 21	13:48:44	16 S	174 W	100	BBBh
Tonga-Fiji region, shocks 300 to 400 km. deep.						
44 x	1937, April 16	03:01:37	21½S	177 W	400	BAAh [12]
47 m	1938, April 25	09:20:16	19 S	176½W	400	CCCh
Kermadec-Fiji region, very deep shocks.						
49 x	1938, April 15	02:59:17	33 S	179 E	580	CCCh
50 i	1938, May 16	01:07:20	27 S	179 E	600	CCBh
50 n	1937, June 19	17:07:27	26½S	178½E	650	BBBh

[5] Huancayo has *P* at 06:36:04, *S* at 06:38:38.
[6] Huancayo has *P* at 05:10:55, *S* at 05:13:36.
[7] Pasadena has pP-P=25 seconds, sP-P=35 seconds.
[8] Huancayo has *P* at 18:37:34, pP at 18:38:12, *S* at 18:41:23.
[9] Aftershock of No. 30 (1935 Dec. 14).
[10] Pasadena has pP-P=02:14, Riverside has pP-P=02:11.
[11] Based on readings of five New Zealand stations and on macroseismic data. Investigated at Wellington. Aftershock: Nov. 1, 23:21:20.
[12] Westland (1938) has a note on this shock.

TABLE 1.—*Data for deep-focus earthquakes—Continued*

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
Kermadec-Fiji region, very deep shocks—*Continued*						
50 t	1937, May 10	15:25:38	26½S	178 E	640	CBBh
52 i	1934, Dec. 12	08:40:49	25 S	178 E	600	BBAh
60 g	1933, Jan. 23	18:14.6	21 S	180	600	CCCe
60 p	1936, Nov. 15	21:50:16	21 S	178 W	540	BBBh
63 m	1938, Mar. 6	16:54:05	19 S	178 W	500	CCCh
64 a	1936, Nov. 26	08:33:32	18½S	178 W	560	BBBh
Loyalty Islands to New Guinea.						
67 e	1935, Aug. 17	01:44:42	22½S	171 E	120	BBBh
67 i	1931, July 21	03:36:22	21 S	170 E	140	AAAb
67 k	1938, June 23	12:55:28	20¾S	169½E	90	BBBh
67 l	1938, May 30	14:29:50	20½S	169½E	70	BBBh
67 m	1938, June 30	16:44:48	20½S	169½E	70	BBBh
67 p	1934, April 11	21:11:57	19½S	169½E	150	ABBhg
69 i	1934, June 3	16:15:40	15½S	168 E	120	BBBhg
69 r	1933, Jan. 1	08:48:39	14¾S	168 E	140	AAAa
76 d	1937, Sept. 15	12:27:32	10½S	161½E	80	AAAh
76 m	1938, Sept. 7	12:58:20	6½S	155 E	160	BAAh
76 p	1934, May 13	09:02:09	5¼S	154 E	100	ABBhg
76 r	1937, May 31	15:31:57	6½S	154 E	100	BBBh
76 s	1937, Aug. 5	14:43:58	5½S	150 E	140	BBBh
76 x	1933, Jan. 15	18:02:02	5½S	147½E	140	BABb
79 m	1934, Mar. 16	14:13:41	5½S	146 E	120	BBAhg
80 a	1935, May 21	06:51:50	6 S	146 E	140	BBBh
80 m	1937, May 12	02:44:55	4½S	144 E	150	BBBh
81 s	1937, April 5	06:56:41	1 S	133 E	90	BBBhn
Solomon and Bismarck Islands, deep shocks.						
86 f	1938, Aug. 31	17:45:13	4 S	151½E	350	BBBh
Sunda Islands, intermediate shocks.						
86 r	1933, Oct. 17	12:23:52	3½S	131 E	130	BCBhg
86 v	1936, Mar. 4	06:30:00	5 S	131 E	150	CCChn
92 f	1936, April 28	13:35:45	6½S	129 E	200	BBCh
92 g	1938, April 26	12:53:35	6½S	129 E	150	CCChn
93 g	1932, Nov. 22	14:51:25	8 S	128 E	180	BBCan
93 r	1932, May 10	14:23:03	7½S	128 E	170	BCCan
95 a	1933, May 21	21:23:50	7 S	127 E	180	BCCb
97 e	1938, Oct. 20	02:19:27	9 S	123 E	90	ABAh
98 a	1936, April 15	06:05:58	9 S	111 E	100	BBBhn
99 e	1938, Jan. 22	15:27:27	7½S	104 E	100	DCDhn
99 m	1932, July 5	10:52:15	6 S	103½E	80	BBBbn
99 o	1938, Aug. 18	09:30:04	4 S	103 E	100	BBAhn
99 q	1938, Aug. 25	01:28:07	5½S	102 E	70	BCBhn
99 t	1938, Jan. 18	04:20:04	4 S	101½E	100	CCCh
99 v	1938, Feb. 2	09:37:03	3 S	100 E	120	CCChn
99 x	1938, July 29	13:06:38	1 S	99 E	80	BCBhn
100 m	1932, June 16	01:18:45	3 N	97½E	80	AAAbn
102 g	1937, Aug. 4	23:35:22	6 N	95½E	120	BBBhn

TABLE 1.—*Data for deep-focus earthquakes—Continued*

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
Banda Sea and Flores Sea, about 400 km. deep.						
104 g	1938, April 4	21:09:03	7 S	127 E	420	BCBhn
Java Sea and Flores Sea, deep shocks.						
107 a	1938, May 8	14:40:35	6 S	124 E	700	CCChn
109 m	1934, June 29	12:34:45	5¾S	121½E	670	ABBhg [13]
110 r	1937, Aug. 11	00:55:54	6¼S	116½E	610	AAAhn
112 a	1938, Feb. 1	18:52:17	6 S	113 E	600	CCChn
116 d	1932, Aug. 11	11:49:33	6 S	110 E	600	CBBcn [14]
Celebes to Mindanao, shocks with depths between 100 km. and 400 km.						
117 m	1932, July 29	20:58:37	½S	124 E	200	ABBan
118 a	1937, May 13	18:47:55	0	123½E	200	BBBh
118 b	1937, Sept. 28	13:17:43	0	123½E	200	BBBhn
Celebes to Mindanao, very deep shocks.						
131 m	1935, Oct. 4	05:15:36	6 N	125 E	500	ABBh
Luzon to Kiushiu.						
133 t	1938, Feb. 5	09:55:10	14 N	124 E	160	BBBh
135 m	1937, Mar. 16	15:45:46	18 N	121 E	100	AABh
135 p	1938, May 23	08:21:53	18 N	119½E	80	ACBh
136 m	1934, April 13	22:03:53	25½N	124½E	250	BBChg
136 t	1932, Dec. 26	21:14:44	26 N	125 E	210	AAAa
138 m	1933, July 18	11:24:54	28½N	129 E	100	BBChg
141 a	1935, Oct. 2	09:27:46	31 N	130 E	120	AABh
141 f	1932, Nov. 17	20:11:40	31 N	130½E	100	AABbk
Marianne Islands–Japan–Kamchatka, intermediate shocks.						
141 n	1932, Nov. 3	19:42:53	16½N	146½E	170	BBBa
141 r	1936, Nov. 12	02:15:58	17½N	147 E	170	BCCh
141 w	1934, Oct. 21	17:53:44	18 N	146 E	210	BAAh
142 i	1935, July 29	04:12:49	19 N	146 E	200	BBAh
142 r	1937, May 5	21:15:39	19¼N	145¾E	190	AAAh
144 a	1935, Dec. 14	12:47:29	21½N	143½E	270	BBBh
144 d	1936, Mar. 31	03:33:10	21½N	143¾E	290	ABBh
144 h	1933, May 21	21:54:05	22 N	145½E	200	CCCc
144 m	1932, Dec. 26	22:31:06	22 N	143½E	280	AAAa
144 r	1932, Sept. 2	12:56:30	23 N	142½E	140	ABBa
147 r	1933, Mar. 18	15:51:32	32¼N	139¾E	170	AAAak
150 d	1934, April 15	10:33:18	34¼N	140½E	70	AAAhg
150 i	1936, Oct. 25	15:30:25	34¾N	140½E	80	AABh
150 m	1937, May 15	12:22:56	35 N	139½E	80	ABBh
150 p	1938, June 5	16:31:38	36 N	140¼E	80	BBBh
150 t	1934, May 30	23:03:38	36¼N	140½E	70	AAAhg

[13] About 4 hours later than No. 109 (June 29, 1934, 08:25:17) with distinctly different epicenter (distance about 2½°) and depth (difference about 50 km.).
[14] Pasadena has F'' at 12:07:37, pP'' at 12:09:45.

TABLE 1.—*Data for deep-focus earthquakes—Continued*

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
Marianne Islands–Japan–Kamchatka, intermediate shocks—*Continued.*						
150 v	1938, Sept. 21	18:52:00	36¼N	141 E	60	AAAh
151 a	1938, May 23	07:18:33	36½N	141 E	70	AAAh
151 e	1938, Feb. 7	14:43:00	36½N	139½E	100	BBBh
151 m	1934, April 6	19:09:36	37¼N	141½E	100	ABBhg
151 r	1937, July 26	19:56:37	38½N	141½E	90	AAAh
152 f	1933, Jan. 3	22:41:04	41 N	138 E	280	BCCa
152 p	1936, June 3	02:55:36	41½N	142 E	80	BBBh
156 k	1938, Oct. 17	15:26:55	44 N	142½E	250	BCAh
158 d	1935, Oct. 2	05:33:03	44 N	146 E	70	AAAh
158 e	1932, Dec. 8	15:17:00	44 N	146 E	100	CCCa
158 m	1934, June 13	01:50:54	44 N	147¾E	90	AAAhg
158 r	1938, Nov. 13	13:13:40	44½N	149½E	70	BAAh
161 m	1934, May 28	05:32:44	45½N	149½E	140	ABBhg
162 m	1937, June 8	18:00:35	46½N	150 E	130	BBBh
168 d	1934, Mar. 18	04:33:15	49¾N	156½E	100	AABhg
168 p	1937, July 15	19:03:26	52 N	159 E	140	BBBh
Aleutian Islands—Alaska, intermediate shocks.						
170 d	1937, Sept. 3	18:48:12	52½N	177½W	80	AAAh
170 g	1939, Feb. 24	14:15:45	53 N	164½W	70	CCAh
170 m	1934, June 18	09:13:50	60½N	151 W	80	AAAhg
Marianne Islands—Manchuria—Kamchatka, deep shocks.						
171 f	1934, Feb. 4	03:10:45	18½N	145 E	570	BBBhg
171 r	1937, May 28	19:56:03	24 N	142 E	530	AAAh
171 s	1937, May 29	02:00:01	24 N	142 E	530	BBBh
171 w	1937, June 12	18:08:10	26 N	141 E	430	BBBh
175 d	1937, Jan. 5	11:09:12	27½N	139 E	500	BABhp
176 f	1933, Feb. 4	06:17:58	28 N	139 E	550	BBBa
176 i	1938, Sept. 21	11:36:20	28 N	140 E	400	CDBh
176 r	1932, Oct. 1	15:08:05	28 N	140½E	450	ABBak
183 g	1935, Feb. 10	18:29:33	29½N	139 E	500	CCCh
183 p	1934, April 19	16:13:29	29¾N	139¼E	430	AAAhg
185 g	1932, July 27	00:30:53	31 N	139 E	380	AABak
185 p	1932, Oct. 14	12:35:57	31½N	139 E	330	ABBak [15]
186 m	1933, Feb. 9	03:56:53	31¾N	138¾E	370	AAAak
187 g	1936, June 25	16:51:52	32½N	137½E	370	BABh
187 r	1933, Mar. 28	23:40:04	32½N	138 E	400	AABa
188 m	1932, Dec. 5	00:19:20	33½N	137 E	420	AABak
191 g	1936, Oct. 26	09:33:32	34¼N	136½E	380	AABh
191 m	1932, May 5	04:11:01	34½N	135½E	380	AAAak
191 r	1932, April 28	03:43:04	34½N	137 E	370	AABak
193 m	1932, July 25	08:24:39	35½N	135½E	360	AAAa
193 n	1937, Nov. 22	04:53:03	35½N	135½E	360	BABh
195 m	1935, April 15	11:15:07	36¼N	137¼E	270	AAAh
197 f	1936, Oct. 19	19:56:00	37 N	135 E	350	BABh
198 g	1935, May 31	08:18:37	38¾N	134 E	450	AAAh
203 f	1935, Mar. 28	23:47:51	43 N	131 E	550	AAAh
203 g	1933, July 14	16:03:35	43 N	131 E	530	BBBhg

[15] Pasadena has *P* at 12:47:43.

TABLE 1.—*Data for deep-focus earthquakes—Continued*

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
Marianne Islands–Manchuria–Kamchatka, deep shocks—*Continued*.						
204 h	1938, Oct. 21	06:46:22	43½N	131 E	550	BCBh
206 m	1933, Sept. 9	05:02:35	44 N	130 E	590	ABAhg
211 m	1937, April 29	20:18:58	46½N	137 E	370	BBBh
213 m	1937, July 21	00:07:37	46 N	143 E	400	BCCh
214 h	1932, Oct. 25	17:02:12	46¾N	144 E	410	AAAa
214 s	1933, Dec. 4	19:33:55	47 N	144 E	360	AAAhg
214 v	1939, April 21	04:29:04	47½N	139¾E	520	AAAh
218 a	1936, Mar. 1	10:21:56	47½N	146½E	440	BBBh
218 m	1933, May 24	04:35:48	47½N	145½E	420	BABa
219 d	1935, July 26	08:03:39	48 N	145½E	480	AAAh [16]
219 m	1935, July 27	10:13:09	48¾N	146½E	490	AAAh [16]
223 m	1933, Jan. 18	17:15:01	51 N	149 E	570	BBBa
Burma.						
229 a	1934, June 2	05:54:29	24½N	95 E	130	AAAgh
229 h	1938, April 14	01:16:35	23½N	95 E	130	AAAh
Hindu Kush.						
247 i	1933, Jan. 20	12:12:12	36½N	70½E	230	BBBa
247 r	1933, May 27	22:41:58	37 N	70½E	230	CCCa
249 h	1935, Feb. 3	02:10:47	36½N	70½E	230	AAAhg
249 m	1935, April 3	11:11:59	36½N	70½E	250	ABBhg
249 t	1935, Oct. 11	04:20:18	36½N	70½E	230	ABBhg
250 b	1937, Oct. 29	07:26:30	36½N	70½E	230	BBBh
250 d	1937, Nov. 14	10:58:12	36½N	70½E	240	AAAh [17]
250 f	1938, Jan. 18	09:29:02	36½N	70½E	250	BBBh
250 g	1938, Jan. 26	10:48:12	36½N	70½E	250	CCCh
250 h	1938, April 6	01:14:30	36½N	70½E	240	CCCh
Western Asia and southeastern Europe.						
252 p	1938, July 13	20:15:17	45.7N	26.7E	150	AABh [18]
253 m	1938, April 13	02:45:46	39.2N	15.2E	270	AAAh

[16] The epicenters of the shocks on 1935, July 26 and July 27 differ clearly by about 1°.
[17] J. Lynch (1938) finds as epicenter 35.6° N., 70.8° E.; origin time, 10:58:14; depth 250 km.
[18] G. Demetrescu (1938) finds as epicenter 45° 43′ N., 26° 45′ E.; origin time, 20:15:13; depth 169 km.

TABLE 2.—*List of shocks found to be normal or slightly deeper*

Letter references are the same as in Table 1.

No.	Date	Time hr.:min.:sec.	Lat. degr.	Long. degr.	Depth km.	Note
S 9 m	1937, Mar. 23	00:44:26	36½S	98 W	50	BBBh
S 17 p	1938, June 9	19:15:11	3½S	126½E	60	AAAh
S 28 m	1932, June 11	17:00:00	13½N	146½E	60	ABBb
S 29 d	1937, Nov. 26	10:45:14	24 N	123 E	70	BBBh
S 29 m	1932, May 28	02:21:19	28 N	132 E	60	AAAb
S 29 p	1937, Jan. 23	08:49:52	33 N	131 E	100	BCCh
S 30 m	1932, Sept. 3	11:58:54	41¼N	143¼E	50	AAAb
S 31 p	1938, July 24	13:12:13	53½N	167 W	50	BCBh

DISCUSSION OF MAPS

Because of the small scale, no world map of epicenters has been drawn for the present paper. There are no significant changes or additions to be made except in the Pacific region. Accordingly, maps have been drafted showing epicenters in the western Pacific area and in the Americas, the latter appearing as insets.

Because of their correlation with the epicenters of intermediate shocks (to be discussed presently), the structural trends of orogenies of late Cretaceous or early Tertiary age have been drawn on these maps. These structural lines are taken from the maps given by A. Born (1933).

Two separate maps have been drawn for intermediate shocks and for deep shocks respectively. In the previous paper the authors took the level separating these two groups at a depth of 250 km.; but in drawing the present maps all shocks at depths of less than 300 km. have been included with the intermediate group. With this change, the epicenters of intermediate and of deep shocks form two clearly distinguishable geographical groups. The distinction between intermediate and deep shocks is as definite as that between normal and intermediate shocks.

Figure 1 shows the epicenters of all intermediate shocks which the authors have located in the Pacific region. Thirty-one shocks in the Hindu Kush, 5 in western Asia and southeastern Europe, 2 in the South Atlantic and 1 in the southeastern Pacific have not been included. As in the previous paper, shocks at depths of less than 150 km. have been indicated by erect crosses, and the deeper intermediate shocks by diagonal crosses. These subgroups tend to separate geographically, but the division is not sharp and is affected by errors of location. The clearest case of this kind is in the central Andes, where a considerable group of shocks at depths of over 200 km. occurs beneath the interior ranges, while shocks of the shallower group occur under the coastal front of the mountains.

A very important new addition is the group of epicenters in central Mexico. Most of these shocks have occurred since January 1937; this suggests that other gaps on the map may be filled in the course of time. It becomes necessary to modify what was stated in the previous paper concerning the association of intermediate earthquakes with oceanic deeps; the Acapulco Deep no longer constitutes an apparent exception. The map strongly suggests that the true correlation is not so much with oceanic deeps as with the structures of which these deeps are a part. Similarly, the frequent occurrence of intermediate shocks along lines of active or recent volcanism must be taken as expressing a connection between these shocks and the tectonic lines along which the volcanism

occurs, rather than any direct causal relation between volcanic activity and intermediate earthquakes. It is noteworthy that there are many more intermediate earthquakes in the volcanic region of Central America than along the structural loop through the West Indies.

The location of numerous new epicenters in South America is largely due to the use of the Huancayo seismographs. The circumstance of having the Huancayo and Pasadena files available together has made possible much more detailed and precise investigation than could have been undertaken otherwise. The maintaining of the Huancayo station by the Carnegie Institution of Washington is a very considerable contribution to seismology.

It is striking that the majority of South American earthquakes, large enough to be recorded at Pasadena or other distant stations in recent years, prove to be at intermediate depths. Even some of those taken as "normal" are at depths of the order of 50 km., with the corresponding *pP*, well recorded; such shocks are not listed in this paper.

It is especially noteworthy that the great and disastrous earthquake in Chile on January 25, 1939, proves to have had a focal depth of not less than 70 km.

As mentioned, the correlation with the Andean group of structures is especially clear. Most striking is the accumulation of epicenters at about 24° S. Here the eastern Andean structures come to an end toward the south, and the corresponding series of epicenters breaks off sharply.

Only one shock (No. 41f) at abnormal depth is mapped in New Zealand. The reports of the New Zealand stations have occasionally noted earthquakes apparently of deep-focus type in this area. Thus far, none of those in New Zealand itself have been large enough to permit a verification from the records of distant stations. A few lying to the north, along the Kermadec-Tonga structural line, have been verified and are mapped in Figure 1. There are also two shocks at short distances from Apia, which very probably are at depths of 150 to 200 km., but for which no precise epicenter can be given; they occurred on June 16, 1932, 23^h, and April 28, 1938, 7^h. These are not shown on the map.

Our knowledge of activity in the New Hebrides region has been extended by the location of a number of shocks at depths exceeding 100 km. in the southern part of the New Hebrides group east of New Caledonia. The new epicenters follow the structures fairly well, but suggest a slightly different trend. A gap still remains between these and the epicenters of the Kermadec-Tonga group, and it is premature to speculate whether this should be filled, and if so, along what line.

The activity extending from the western Solomon Islands into New Guinea is now better established, but a striking gap still remains between

these and the shocks of the Banda Sea region. This is not significantly affected by the location of shock No. 81s, at 1° S. 133° E. in the western projection of New Guinea. This epicenter, however, has important struc-

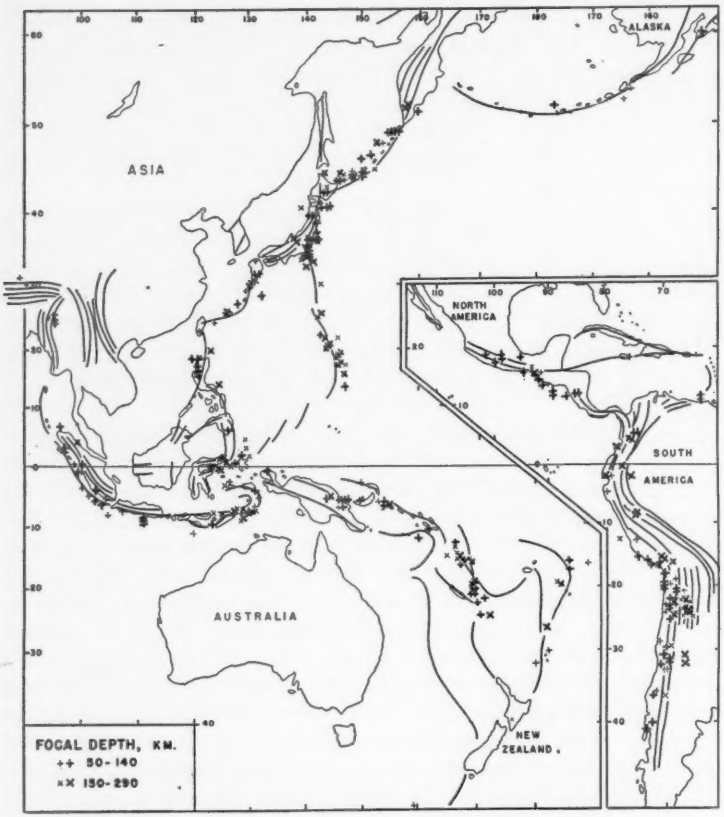

FIGURE 1.—*Epicenters of intermediate shocks in the Pacific region*
Smaller symbols indicate less accurate determination.

tural implications, and it is to be hoped that future shocks will provide confirmation.

The structural arc extending from the Banda Sea to Java and Sumatra is now better indicated, and it is reasonable to expect that in the course of time it will be completely outlined by the epicenters of intermediate shocks. There still remains a striking gap between Sumatra and Burma.

There are very few new epicenters in the Philippine region; the relation to the structures is incompletely indicated, but appears to be the same as elsewhere.

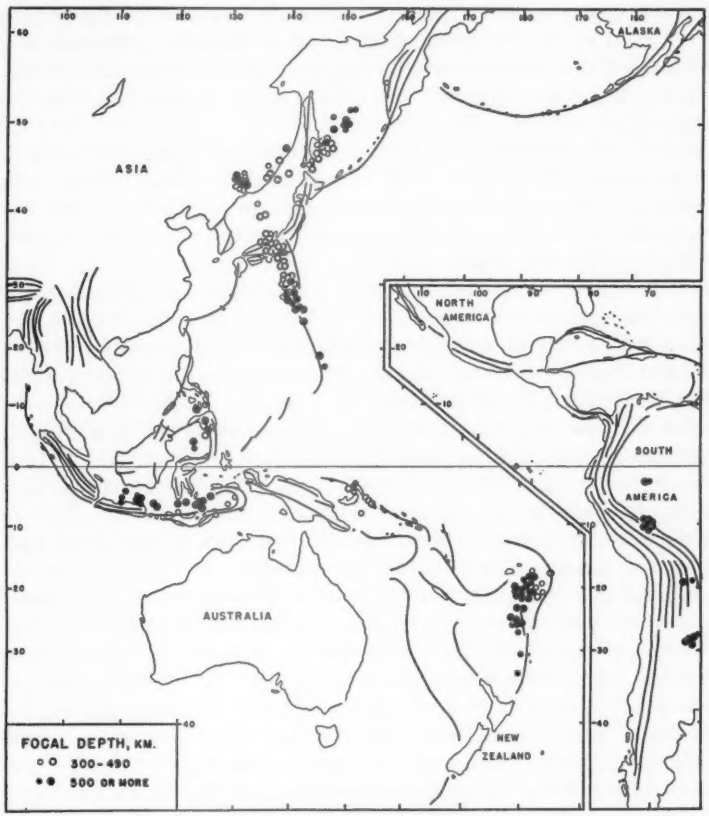

FIGURE 2.—*Epicenters of deep shocks*
Smaller symbols indicate less accurate determination.

The new data serve to bring out very clearly the line of epicenters extending through the Ryukyu Islands to Kyushyu. The gap between these epicenters and the intermediate shocks in central Honshu is very striking and exceptionally well established, as comparatively small shocks can be used with confidence in this region. The major line of intermediate shocks in the Japanese region extends continuously from

the Marianne Islands by way of eastern Honshu to the Kurile Islands and Kamchatka.

Four intermediate shocks have been located along the Aleutian-Alaskan arc. Two of these are well determined, with depth about 80 km.; the third is assigned a depth of 50 km., but may be shallower and classifiable as a normal shock; the depth of the fourth is tentatively determined as 70 km., but the information is incomplete. Other intermediate shocks in this region are suspected, but the data are usually insufficient or suggest depths of the order of 50 km.

It is a striking fact that a large and conspicuous gap remains in the circum-Pacific distribution of intermediate earthquakes. This is the North American region between Alaska and central Mexico. There is no doubt about this, as even a comparatively small intermediate shock could be recognized and located in this region.

Since publication of the previous paper, activity in the Hindu Kush region has continued. The shock of November 14, 1937 (No. 250 d) is one of the largest known to have originated from this source, and was recorded in detail at many stations. The epicenters and depths continue to be confined within narrow limits.

The new list contains two of the rare intermediate shocks of the European region. The Rumanian shock of July 13, 1938 (No. 252 p) is close to the shock of Nov. 1, 1929 (No. 253), and has nearly the same depth. Of more significance is the shock of April 13, 1938 (No. 253 m), off the coast of Italy, north of Stromboli. This is the deepest shock found to have occurred outside of the Pacific region.

For the Pacific region, correlation with Tertiary structures, particularly where there is volcanic activity, remains very well marked. It is possible to suggest where shocks may be expected in the future at locations not thus far represented on the map. On the other hand, the location of intermediate shocks may eventually furnish material for a revision of the structural lines in regions where other data are inadequate.

Certain conspicuous gaps persist, and in some cases there is reason for expecting this condition to be permanent. There are several regions of Tertiary structure in which normal shocks are large or frequent, without any accompanying intermediate shocks; among these are northern Mexico and central New Guinea.

The absence of intermediate shocks in the southwestern half of Honshu has been noted. Recent volcanic activity is also low in this part of the island. It is striking that no intermediate shocks have been found to be associated with the volcanoes of Sakhalin (Karafuto). In general, there is indication that in the Japanese region the lines of epicenters of intermediate shocks diverge from the pattern of Tertiary structures.

Epicenters for intermediate shocks outside the Pacific region are still too few to permit much generalization. Those in the South Atlantic (Southern Antilles) probably actually belong to the Pacific group. The remainder are associated with the Alpide belt of structures. The persistence of the Hindu Kush shocks from so small a focal region is so unique that it suggests the existence of a special condition there.

Figure 2 shows the epicenters of all shocks at depths of 300 kilometers or more, which the writers have been able to locate to date. All these lie in the Pacific region. Where the epicenters are sufficiently numerous, they are found to lie along rather narrow belts, regardless of depth. It must be emphasized that there is a clear geographical distinction between intermediate shocks at about 250 km. and the true deep shocks near 300 km. The epicenters are usually widely separated except in limited regions (such as central Japan) where a line of intermediate epicenters and a belt of deep shocks approach each other.

In contrast with the large number of additional intermediate shocks in South America, the authors have listed only two new deep shocks there, both from epicenters already mapped. They have located no shock at great depth in South America since that of January 14, 1936 (No. 40). Activity from deep sources in that region seems to consist chiefly of large shocks at long intervals; thus the authors' catalogue shows large shocks on August 4, 1930, August 29, 1933, and December 14, 1935, with no others between. Small shocks of this group might have escaped notice in earlier years but in recent years this is less likely. It is still doubtful whether or not the four separate regions marked by epicenters of these shocks form part of a continuous active belt.

In the Kermadec-Tonga region the additional epicenters do not materially change the situation. Shock No. 49 x (April 15, 1938) is the most southerly shock of this entire group. It is reported as having been felt at Wellington, which is about 1000 kilometers distant from the epicenter. Shock No. 48 (Jan. 1, 1935) occupies an anomalous position northeast of all other epicenters indicated on the map. The depth of this focus is only 300 km., and it may well be that it should have been included with the intermediate shocks.

In the Bismarck Islands there is only one new epicenter, and six in the Banda Sea, Flores Sea, and Java Sea, none of which contribute any essentially new information. In the Philippines there is also only one new deep shock, in southern Mindanao.

Over 30 new epicenters have been added in the Japanese active belts. Depths of 500 km. and over are better represented than before. One of these (No. 171 f, Feb. 4, 1934) is near the southern end of the known active belt, in the Marianne Islands, and has a depth of 570 km. The

continuity of the transverse active belt across the Japan Sea is further supported by No. 198g (May 31, 1935). The shock of April 21, 1939 (No. 214v) is from a new isolated epicenter near the Siberian coast.

The chief conclusion to be drawn from this map is that the truly deep shocks occur exclusively under regions of continental structure

TABLE 3.—*Number of shocks listed at various depths*

Region	Depth in km. (Range ± 25 km.)												
	100	150	200	250	300	350	400	450	500	550	600	650	700
Mexico, Central America	10	3	1	....	....	....	....	....	....	....	....	....	....
South America	26	20	15	5	1	....	....	....	....	....	4	11	....
Fiji-Kermadec	5	2	2	....	2	1	4	....	4	10	7	3	1
Loyalty Is.-New Guinea	14	12	6	1	1	2	3	1	....	....	....	....	....
Sunda Is.	17	8	9	1	....	1	2	1	1	....	9	2	4
Celebes to Mindanao	....	3	8	2	3	....	1	....	2	....	2	1	....
Luzon to Kiushiu	5	4	3	3	....	....	....	....	....	....	....	....	....
Japanese Is.	21	20	10	4	11	24	17	15	10	14	3	1	....
Hindu Kush	1	....	9	21	....	....	....	....	....	....	....	....	....
Others	7	8	1	1	....	....	....	....	....	....	....	....	....
Total	106	80	64	38	18	28	27	17	17	24	25	18	5

surrounding the Pacific Basin. There is no generally applicable correlation with known surface structures of any age. It is difficult to avoid the impression that the deep shocks of the Java Sea, Flores Sea, and Banda Sea are associated with the structural loop south and east of them, and that those of the Kermadec-Tonga region are associated with the structural loop east and north of them. There may be a somewhat similar structural relation in South America, but information is still incomplete. On the other hand, there is no evident correlation of this type in the Japanese area, particularly to the north and west.

STATISTICAL RESULTS

Table 3 shows the distribution in depth of all shocks found to be deeper than 75 km. Each column includes shocks at depths within a range of ±25 km. from the mean depth heading the column. The preponderance of shocks at the shallower depths is more conspicuous than in the previous catalogue (1938). Forty-four shocks at depths of from 50 to 75 kilometers have been omitted from this table, as their inclusion is more or less accidental and might be statistically misleading. It is now clearly established that, for the earth as a whole, the majority of shocks which can be investigated are at normal depth or very slightly

deeper, and that the number of shocks falls off rapidly with increasing depth down to the limit of intermediate shocks at about 300 km. The true deep-focus shocks show no clear preference for any depth throughout their range from 300 km. to about 700 km. Shocks at certain depths continue to be characteristic of certain regions, as will readily appear from the table and maps.

In the previous paper it was mentioned that of 58 listed deep shocks (excluding intermediate shocks) in the Japanese area, 41 occurred in the first half of the year and 17 in the second half. The present paper adds 21 and 17, respectively, to these figures, raising the totals to 62 for the first half and 34 for the second half. The intermediate shocks of the same region do not show this annual distribution; the corresponding totals are 32 for the first half and 29 for the second half. The deep shocks of the Kermadec-Tonga region are the only other group which shows a similar result; the totals are 22 for the first half and 12 for the second half of the year. The total of deep shocks in all other regions gives 23 for the first half and 26 for the second half.

SUMMARY

This paper is in continuation of a previous publication (1938). The writers distinguish (1) shallow shocks, at depths not exceeding about 50 kilometers; (2) intermediate shocks, at depths from about 50 to 300 kilometers; (3) deep shocks. Separate maps are drawn for intermediate and for deep shocks, and new or revised determinations are presented and discussed. Previous conclusions as to mechanism and origin remain unmodified.

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RELICT FRACTURE CLEAVAGE IN PARASCHISTS

BY J. T. STARK AND J. H. MAXSON

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ABSTRACT

To the commonly known criteria suggesting that metamorphosed rocks were originally sedimentary may be added relict fracture cleavage, when confined to certain thin beds. Thin beds in the Archean Vishnu series of the Grand Canyon possess cross structures at high angles to their bounding surfaces. These transverse planes presumably were developed in incompetent layers during crustal deformation. Later they became the loci of biotite crystallization and thus were preserved. Close folding of parts of the Vishnu series on Vishnu Creek indicated by reversals in attitude of fracture cleavages is corroborated by the testimony of drag folds.

INTRODUCTION

Geologists whose work deals with old rocks, especially the metamorphic highly altered crystalline pre-Cambrian complex, often face the problem of determining what these altered rocks were originally. Heat, pressure, shearing movements, and recrystallization may change igneous and sedimentary formations so greatly that little of their primary characteristics remain, and in many petrographic problems the question is not what to expect in the metamorphism of a particular rock but rather what to think the metamorphic rock was before it was altered. Any addition to the criteria suggesting the origin of metamorphic rocks is of interest to the geologist.

Aside from actual gradation of schists into sedimentary rocks and certain diagnostic chemical compositions, the common distinction of metamorphic rocks of sedimentary origin from those of igneous origin has been the preservation of such original structures as stratification,

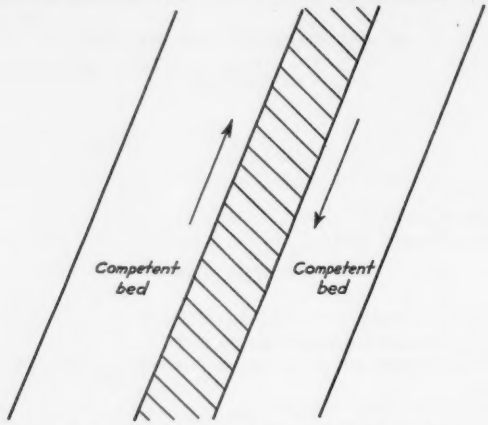

FIGURE 1.—*Fracture cleavage developed in an incompetent bed*

cross bedding, detrital fragments, fossils, ripple marks, and such secondary structures as concretions. Another type of criterion might be provided by preservation of structures where these may be shown to be characteristic of the deformation of sedimentary rocks. The occurrence of a structure, which has been interpreted as relict fracture cleavage, in the Archean paraschists of the Vishnu series in the Grand Canyon forms the subject of this paper.

The structure in question was discovered in Vishnu Canyon by the Carnegie Institution-California Institute of Technology Colorado River Expedition of 1937. In previous studies the presence of sedimentary structures had been shown by Campbell and Maxson (1933, p. 806-809; 1934, p. 298-303). Chemical analyses of certain bands suggest by high silica content (Campbell and Maxson, 1935, p. 324) the sedimentary origin of the schist. The term Vishnu series has therefore been restricted to include only the paraschists.

VISHNU SERIES

The upper Granite Gorge between Mineral and Zoroaster canyons contains a magnificent exposure of the schists. Between Mineral and Cotton-

SLAB SHOWING RELICT FRACTURE CLEAVAGE
Mouth of Grapevine Canyon.

FIGURE 1. CROSS BEDDING IN QUARTZ-SERICITE SCHISTS
Lone Tree Canyon.

FIGURE 2. SLAB SHOWING RELICT FRACTURE CLEAVAGE
Faintly outlined, mouth of Vishnu Canyon.

CROSS BEDDING AND RELICT FRACTURE CLEAVAGE

wood canyons the schists have been complexly intruded and granitized. Between Vishnu and Zoroaster canyons the schists are less affected by the invading granites and pegmatites. In general, the schists are light gray and are composed chiefly of quartz, feldspar, biotite, and sericite. Near the mouth of Clear Creek there are interbanded sericitic quartzites

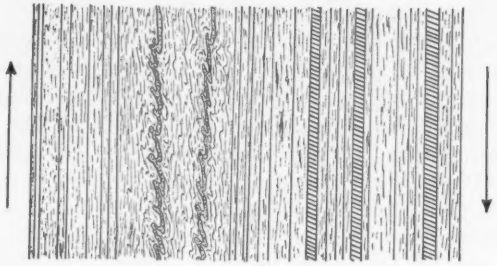

FIGURE 2.—*Relationship of fracture cleavage and drag folds*

and amphibolites. The schists all exhibit regular and persistent banding, individual bands ranging in width from a few inches to several feet. The banding is due to varying mineralogical composition which is believed to be inherited from original stratification. The regional strike of the schistosity averages N. 70° E. The dips are almost universally steep, departing in only a few outcrops more than 15 or 20 degrees from vertical. The majority of measurements show a northward dip, but reversals are not infrequent. The strike and dip of the regional schistosity appear everywhere to be parallel to the banding, except as mentioned later.

RELICT STRUCTURES

In the walls of Vishnu Canyon which cut across the regional schistosity, some bands were observed to have regular cross structures. These are formed of biotite plates in the transverse planes at high angles to the bounding planes of the adjacent bands (Pl. 1). The transverse structures are not distributed haphazardly in the schist but are confined to certain definite bands. When found in one, they are present throughout its exposed surface. Adjacent bands have none. Although at first suggestive of cross bedding, their regularity, high angle of inclination, and lack of curvature or feathering near the boundary planes oppose this explanation. Figure 1 of Plate 2 represents for purposes of comparison a structure in the schist which has been interpreted as cross bedding. On the other hand, the transverse structures in Plate 1 closely resemble fracture

cleavage formed in sedimentary rocks subsequent to lithification. In a characteristic specimen, a 2.5 cm. band with cross structures is exposed. These structures make angles as high as 85 degrees with the bounding bands and are spaced approximately 6 mm. apart. The band containing the cross structures is highly sericitic as contrasted with its feldspathic and quartzose neighbors. It may have been a shaly bed between two sandstone strata and therefore relatively incompetent.

If the cross structures were due to development and concentration of biotite along fracture cleavages, one might expect, from known angular relations of fracture cleavage to bedding on the limbs of folds, to establish whether or not the schists were closely folded by the presence or absence of reversals in attitude of the structures. Such reversals were found in traverses along Vishnu Creek and adjacent portions of the canyon of the Colorado River, indicating clearly that if the cross structures are following directions of fracture cleavage, then extremely close folding has affected the Vishnu schists. According to Leith (1923, p. 155),

> "The fracture cleavage, while at an angle to flow cleavage, has qualitatively the same angular relations to bedding as flow cleavage, and like the flow cleavage it may be used as evidence of differential movement between beds. . . ."

The differential movement of beds during deformation apparently produced drag folds in the more incompetent layers. These structures are obscured in most places by the regional schistosity but are emphasized where concordant bodies of quartz and pegmatite intruded the schists subsequent to the folding. In the lower reaches of the Vishnu Creek section igneous intrusions are relatively scarce, but, as the valley head is approached, many thin lenses, commonly less than a foot long, have the form of drag folds. A similar occurrence of structures, outlined by lit-par-lit stringers of pegmatite, was noted by the senior author (Stark, 1932, p. 477) in the pre-Cambrian schists of central Colorado, where the accordance of drag fold axes permitted mapping of a regional structure throughout the entire 80-mile length of the Sawatch Range that would have been obscured but for the color contrast between the intruded material and the coarsely recrystallized schists. In the Vishnu Creek section, as in Colorado, the concordant intrusions reveal a structure that in most places is obliterated by the regional schistosity. If the cross structures are relict fracture cleavage, the direction of movement indicated by them (Fig. 1) should accord with that indicated by drag folds. A careful traverse of the Vishnu Creek section showed that the directions of movement revealed by drag folds check in all places with direction of movement suggested by the attitude of the cross structures, assuming that they represent fracture cleavage (Fig. 2). Convincing corroboration of

close folding was found in vertical cross sections of several small anticlines and synclines. These are graphically outlined by abundant quartz and pegmatite stringers in the form of drag folds on limbs and across crests and troughs of folds. Bands with biotite cross structures follow

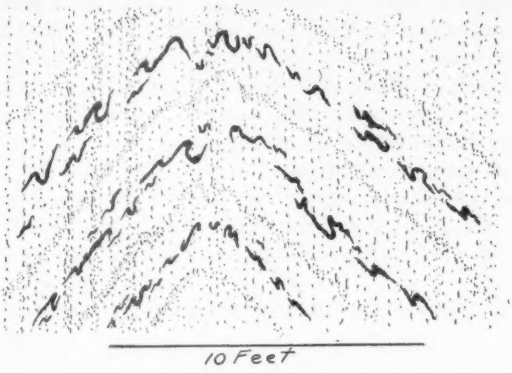

FIGURE 3.—*Sketch showing vertical section across small anticline in paraschists, Vishnu Canyon*

Drag folds of quartz and pegmatite (in black) outline the old bedding. Location of relict fracture cleavage is indicated by shaded areas transverse to the regional schistosity.

the folding in exactly the same manner as bedding with fracture cleavage (Fig. 3).

In some exposures the development of the biotite concentrations is pronounced, and in other places the flow cleavage of the regional schistosity has almost completely obliterated the earlier structure. In many places the cross structures are so faintly suggested that they might be easily overlooked if not sought for specifically. They are recognized in many sections of the Vishnu schists in the Granite Gorge (Pl. 2, fig. 2).

The biotite concentrations responsible for the transverse pattern are best seen when the outcrops are viewed from a slight distance. In many exposures the pattern is not apparent when viewed at close range or in hand specimens. However, once the pattern was recognized and looked for in new exposures it became readily apparent in many outcrops. The steep walls of Vishnu Canyon that had seemed only regular and nearly vertically banded schists with a pronounced schistosity parallel only to the bedding, when revisited with the idea of cross laminations in mind, showed in many places the relict structures faintly outlined by biotite flakes.

IMPLICATIONS AND CONCLUSIONS

If the cross structures are initiated by pre-existing fracture cleavage—and the many folds showing drag folds in accordance with this interpretation leave but small doubt that this is the case—would fracture cleavage on so large a scale be expected in any but stratified sediments? Banding occurs in igneous formations, and shear zones may have fracture cleavage developed in them. But where exposures extend over many miles and especially where they show continuous but thin bands which average less than 6 inches in width—commonly very much less, with fracture cleavage-like structures so widely developed—it seems highly improbable that such formations were originally anything but bedded sediments. Fracture cleavage in thin beds, developed on such a scale, may be added to the list of structures which, when partly preserved, suggest the original character of completely recrystallized and metamorphosed formations.

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DOLOMITIZATION IN GLACIO-LACUSTRINE SILTS OF LAKE AGASSIZ

BY G. D. SHERMAN AND G. A. THIEL

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ABSTRACT

Fine-grained lacustrine silts from 25 to 75 feet thick cover hundreds of square miles on the floor of glacial Lake Agassiz. In numerous localities an enriched dolomitic zone from 1 to 10 feet thick occurs in the silts at from 1 to 3 feet below the surface. The modes of occurrence of the enriched zones indicate that the dolomitic bodies are not simply a product of negative enrichment formed by the leaching of calcium carbonate. Most of the dolomitization has taken place where the water table is near the surface or where the natural drainage is poor. The ground waters contain a high percentage of sodium and of magnesium salts. Gypsum nodules and crystals are found with and below the dolomite. This association suggests a replacement of calcium by magnesium and the precipitation of calcium as gypsum.

INTRODUCTION AND SCOPE OF INVESTIGATION

When the Late Wisconsin ice sheet receded northward from the Continental Divide that separates the drainage basins of the Red River of

the North and the Mississippi River, the melt waters accumulated along the south and west margins of the ice to form the temporary glacial Lake Agassiz. When at its maximum extent, Lake Agassiz was about 700 miles long and 250 miles wide and covered parts of what are now south-

FIGURE 1.—*Map showing extent of glacial Lake Agassiz during its highest water stage.*

ern Manitoba, eastern North Dakota, northwestern Minnesota, and part of northeastern South Dakota (Fig. 1).

The sediments deposited in the glacial lake include (1) Ridges of sand and gravel washed up by wave action along shore, (2) near-shore sand plains, and (3) off-shore silts and clays. A series of beach ridges were formed at successively lower elevations as the level of the lake declined. Each ridge is from 10 to 20 feet high and generally is steeper on the lake side. The near-shore sands, commonly 10 to 15 feet thick, border the beach ridges and form a sloping plain from 5 to 10 miles wide. On the lake side the sands grade horizontally into silts and clays which cover the broad expanse of the lake bottom. These fine-grained sediments are from 25 to 75 feet thick. Locally in northwestern Minnesota they are more than 100 feet thick.

Since the lacustrine sediment of Lake Agassiz represents outwash from the calcareous Late Wisconsin drift, the silts and clays are rich also in calcite and dolomite. Furthermore, analyses show that the calcareous

till carries from 30 to 80 per cent of its carbonate in the form of dolomite. With such parent material a high dolomite content might be expected in the lake sediments. Investigations on the subsoil in the region of Crookston, Minnesota, indicated, however, that an enriched dolomitic zone occurred in the silts at varying depths below the surface.

The purpose of the present investigation was to determine, if possible, the extent of the dolomitization and its relation to topography, drainage, position of the water table, and to the structure, texture, and composition of the sediments.

METHODS OF STUDY

FIELD

An extensive drainage and highway construction program in northwestern Minnesota has resulted in the digging of hundreds of miles of ditches in the lacustrine sediments of Lake Agassiz. Since the most intensive dolomitization in the sediments is near the surface, it seemed feasible to use the banks of road ditches to measure and sample profiles in the silts. More than 700 miles of road and other ditches were examined. For delimiting dolomitized areas, samples were obtained by trenching or by hand drilling. Eleven trenches were dug, 320 7-inch holes were drilled, and more than 3000 borings were made with soil augers. Detailed studies were made of more than 30 profiles in which a dolomitic zone had been found, and from these profiles 205 samples were collected for laboratory study. In the dolomitic zones the samples were taken at vertical intervals of 2 inches. Where a profile was not sampled in detail, the vertical limits of the dolomite were determined by borings and with acid treatment in the field. The thickness of the dolomitic zones was measured at short horizontal intervals. One channel sample was then taken across the dolomite layer, and one representing an equal thickness above the dolomite, and another below it.

From laboratory tests on the silts, it was noted that the samples showed strong effervescence with cold dilute HCl if no more than 87 per cent of the total carbonate was in the form of dolomite. If over 87 per cent of the total carbonate was dolomite, there was feeble or irregular effervescence with cold acid and violent effervescence with hot acid. This reaction provided a method for detecting the outlines of the dolomitized zones in the field and thereby decreased the number of samples that were required to delimit the enriched zones and lenses. Each profile was cleaned to expose fresh, undisturbed silt before the acid test was applied. The contacts shown in Figures 4 and 5 were determined by this method.

In sampling with soil augers, two bits were used—one with a diameter

of 1⅜ inches and one 2 inches in diameter. The two were used in order to secure samples uncontaminated with material from nearer the surface. Holes were bored with the larger bit until a layer was reached that

FIGURE 2.—*Map of northwestern Minnesota*
Showing location of Averill-Glyndon region in the Lake Agassiz basin.

effervesced with hot dilute HCl but not with cold acid. After the hole had been well cleaned with the same auger, a sample was taken from the bottom with the smaller auger. The sample was then tested with cold acid, and if there was no effervescence the hole was deepened with the larger bit. Thus, by using first the one and then the other auger, the sampling and testing with hot and cold acids was continued to whatever depth was necessary in order to reach the calcareous clays and silts below the dolomitized layer.

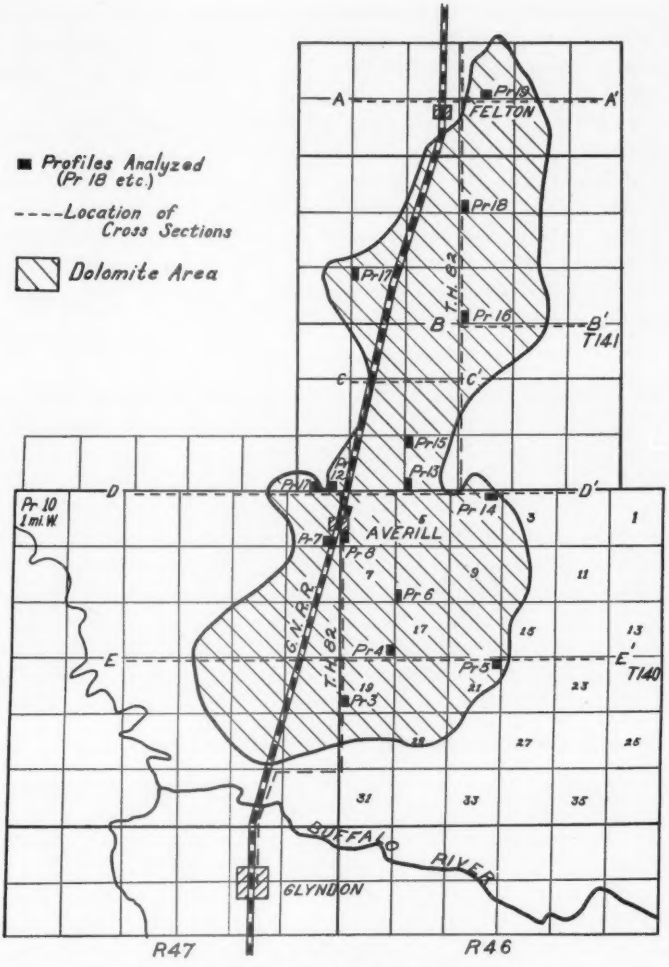

FIGURE 3.—*Map of the Averill-Glyndon region*
Showing geographic extent of one of the dolomitized areas.

DETERMINATION OF CARBONATES

The carbon dioxide content was determined by the absorption tower method. Then a weighed quantity of silt was heated with the volume 0.5N HCl required to decompose the amount of carbonates corresponding to the previously determined CO_2 content. The CaO in the resulting

solution was determined, the $CaCO_3$ computed, and the $MgCO_3$ obtained by difference. In computing the dolomite, it was assumed that all of the magnesium carbonate is combined wtih an equivalent amount of calcium carbonate.

The question may be raised as to whether sufficient ferrous carbonate was present to influence appreciably the computed magnesium content of the dolomitized sediments. Therefore, heavy carbonates were separated from 12 samples by means of heavy liquids, and their ferrous iron content determined. The maximum amount of iron that was found corresponded to 0.85 per cent of ferrous carbonate. In most of the samples the quantity was less than 0.05 per cent. Such small amounts were ignored in making the computations for magnesia by difference.

MICROCHEMICAL TESTS

Various staining and microchemical methods to distinguish calcite and dolomite have been proposed. The writers investigated several recommended methods and concluded that the ferric chloride method was most reliable (Keller, 1937, p. 950). The carbonate minerals were separated from the grains of quartz and feldspar by centrifuging with heavy liquids. The carbonate fractions were then immersed for 5 to 10 seconds in a 2½ per cent solution of ferric chloride. The mineral grains were then washed in water and immersed for 10 seconds in ammonium sulphide $(NH_4)_2S$ which had been saturated with H_2S. This was followed by a second water rinse to remove excess chemicals. With this treatment the calcite grains are stained black, whereas the dolomite retains its original light color (Fig. 3; Pl. 1).

GEOGRAPHIC DISTRIBUTION OF DOLOMITIC SILTS

The most extensive and the thickest layer of dolomitic silt was found in Clay County, Minnesota, about 3 miles north of Glyndon (Figs. 2 and 3). It extends northeastward approximately 13 miles to about 1 mile beyond the village of Felton. The width of the area varies from 1 to 5 miles. By sampling at short horizontal intervals it was found that the dolomitic layer apparently is continuous under the entire area. Smaller areas were found in Wilkin, Norman, and Polk counties. In Polk County a small area on the farm of the Agricultural Experiment Station was sampled in great detail.

ANALYTICAL DATA

The distribution of carbonates and the proportion of carbonates in the form of dolomite are tabulated below. The samples with more than 87 per cent of the carbonate in the form of dolomite represent the enriched zones.

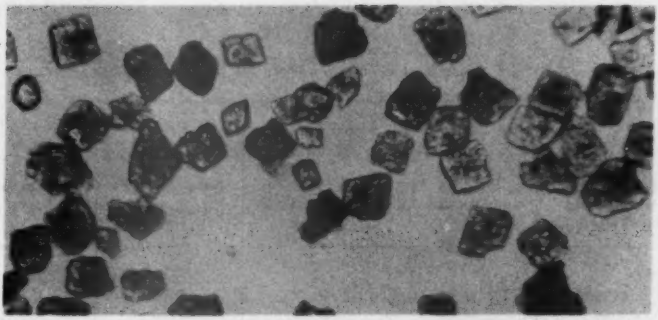

FIGURE 1. CARBONATE GRAINS FROM DOLOMITIZED SILTS OF GLACIAL LAKE AGASSIZ
Stained with ferric chloride. Only a few grains of calcite (black) remain. X 120.

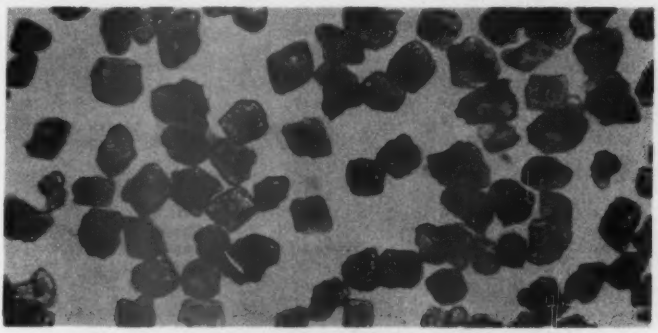

FIGURE 2. CALCAREOUS CARBONATE GRAINS NOT DOLOMITIZED
From below zone of dolomitization. Stained with ferric chloride. Most of the grains are blackened. X 120.

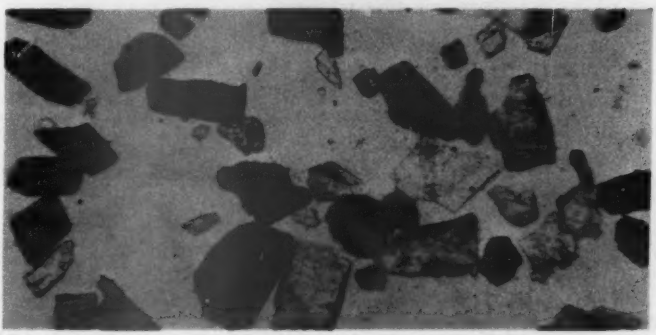

FIGURE 3. FRAGMENTS OF PURE CALCITE (BLACK) AND DOLOMITE (LIGHT)
Stained by ferric chloride method. X 100.

PHOTOMICROGRAPHS OF CARBONATE GRAINS FROM LACUSTRINE SILTS

TABLE 1.—*Distribution of carbonates in the dolomitized silts shown in cross section E-E, Figure 4*

Location of the profiles are shown in Figure 3.

Sample number	Depth in inches	Distribution of carbonates (Per cent)		
		Total carbonates	Total dolomite	Proportion of carbonates in form of dolomite
(Profile No. 3)				
14	22–29	27.81	24.47	88.0
15	29–36	17.70	16.44	92.9
16	36–43	17.92	15.81	88.2
(Profile No. 4)				
17	0–10	21.42	11.59	54.1
18	10–16	31.49	14.64	46.1
19	16–22	31.39	15.63	49.8
20	22–26	22.09	18.75	84.9
21	26–28	22.05	19.25	87.3
22	28–30	21.80	19.90	91.3
23	30–32	23.37	21.69	92.7
24	32–34	24.25	22.79	94.0
25	34–36	25.21	23.44	93.0
26	36–38	24.17	22.69	93.9
27	38–40	22.41	21.74	97.0
28	40–48	22.57	20.97	92.9
29	48–60	19.00	15.88	83.6
30	60–72	25.94	22.18	85.5
31	85–96	29.15	27.02	92.7
32	108–120	28.02	25.22	90.0

Lithology of Profiles

Profile No. 3. About 3½ miles south of Averill in a ditch along the west side of Highway No. 82. Highly dolomitic zone of very fine tan sand, distinguished sharply from grayish sand above and grayish-green lacustrine clay below.

Sample No.

14 Light, grayish-tan, fine, sandy loam.
15 Light-tan, fine, sand mottled with black and rusty stains.
16 Tan, clayey silt.

Profile No. 4. At the SE. corner, sec. 18, T. 140 N., R. 46 W.

Sample No.

17 Grayish-black, heavy, silt loam with numerous white specks.
18 Grayish, silt loam with scattered white specks. Lighter in color than surface layer.
19 Grayish sandy silt.
20-21 Grayish fine sand ranging from mouse gray to grayish tan.
22-26 Light-tan, fine sand with many small black and rusty stains. Color distinguishes it from layers above and below.
27-29 Gray, fine sand with large black and rusty stains, occurring as pockets and lenses.
30 Gray, sandy clay, with many rust stains. Some continuous across beds.
31-32 Gray, very smooth silty clay with many iron concretions and large nodules of gypsum crystals.

TABLE 2.—*Distribution of carbonates in the dolomitized silts shown in cross section D-D', Figure 4*

Locations of the profiles are shown in Figure 3.

Sample number	Depth in inches	Distribution of carbonates (Per cent)		
		Total carbonates	Total dolomite	Proportion of carbonates in form of dolomite
(Profile No. 9)				
49	15–18	37.13	17.60	47.4
50	18–21	30.65	28.81	94.0
51	21–24	27.69	25.83	93.3
52	25–44	27.19	25.50	93.8
53	45–56	22.00	18.38	83.8
(Profile No. 11)				
72	11–21	30.85	26.78	86.8
73	21–31	33.13	30.81	93.0
74	31–41	21.50	18.21	84.7
(Profile No. 12)				
75	0–12	51.82	16.27	31.4
76	12–18	30.32	20.34	67.1
77	18–23	30.64	28.13	91.8
78	23–28	28.70	20.81	72.5
79	30–36	23.46	17.74	75.6
(Profile No. 13)				
87	16–26	23.96	21.01	87.7
88	26–34	28.62	26.76	93.5
89	34–32	30.20	28.27	93.6
90	90–102	36.27	32.54	89.7
(Profile No. 14)				
91	0–9	35.36	7.96	22.5
92	9–18	29.50	7.67	26.0
93	18–24	10.81	8.98	83.1
94	24–29	12.84	12.04	93.8
95	29–34	17.13	14.42	84.2

Lithology of Profiles

Profile No. 9 has a dolomitized Zone 26 inches thick, the upper part of which is a very fine, buff sand, mottled with iron stains. Lower 3 inches is a silty clay which grades downward into a highly calcitic clay. The sediment above the dolomite is a light, grayish, silty loam.

Profile No. 11 has a dolomitized zone 10 inches thick. It extends from 22 to 31 inches below the surface and is underlain by calcitic lacustrine clay. The sediment over the dolomite is a gray, silty loam.

Profile No. 12 contains a thin layer of dolomite in very fine, buff sand. It is overlain by a grayish, silty loam and grades downward into calcitic lacustrine clay.

Profile No. 13 shows a dolomitized zone extending from 26 inches to 102 inches below the surface. The upper part of the dolomitic concentration occurs in a mottled, gray, fine sand, but from 72 to 90 inches below the surface it is found in a silty clay that contains many small crystals of gypsum.

Profile No. 14 is located at the eastern edge of the dolomitized area. Five sites 25 paces apart were sampled to determine whether there is a gradual change in the amount of dolomite or a sudden decrease in enrichment. The data in Table 2 are from the one farthest to the west. A thin zone of dolomitic enrichment was found at the same depth in four of the five sites. At the most easterly site the sediment showed a marked increase in size of grain. In this coarser sand the amount of dolomite showed a sudden drop to 73 per cent.

TABLE 3.—*Distribution of carbonates*

In profile No. 10 in NW. ¼ sec. 1, T. 140 N., R. 48 W., in bank of ditch east of Buffalo River.[1]

Sample number	Depth in inches	Distribution of carbonates (Per cent)		
		Total carbonates	Total dolomite	Proportion of carbonates in form of dolomite
54	0–12	4.56	2.89	63.4
55	12–24	12.48	7.08	56.7
56	24–36	16.33	9.65	59.1
57	36–48	28.45	17.15	60.3
58	48–60	18.05	14.02	77.6
59	60–72	18.92	14.76	78.0
60	72–84	12.74	9.48	74.4
61	84–96	12.14	8.91	73.4
62	96–108	14.32	11.41	79.6
63	108–120	17.41	14.15	81.2
64	120–132	16.22	14.23	81.5
71	124–125	35.66	32.08	90.0
65	132–144	14.58	11.67	80.0
66	144–156	18.05	16.31	90.4
67	156–168	34.54	31.12	90.1
68	168–180	23.39	20.41	87.2
69	180–192	36.99	33.88	91.6
70	192–204	36.97	34.16	92.4

Lithology

Profile No. 10. The section is exposed in the bank of a ditch where planation has exposed a fresh wall 15 feet above the level of the water in the ditch. The upper 6 feet is composed of a uniformly light-colored clay. This is underlain by 9 feet of typically varved lacustrine clays and silts. The varved sediments rest on sandy loam mottled with brown and black stains.

[1] Profile 10 is located to the west of the area shown in Figure 3. *See* notation at west margin of map.

TABLE 4.—*Distribution of carbonates in the dolomitized silts*

At the southwest corner of sec. 17, T. 132 N., R. 45 W., in Wilkins County, Minn.

Sample number	Depth in inches	Distribution of carbonates (Per cent)		
		Total carbonates	Total dolomite	Proportion of carbonates in form of dolomite
1	6–16	4.57	0.33	7
2	16–18	0.86	0.24	28
3	18–20	1.76	1.43	81
4	20–22	3.19	2.93	93
5	22–24	3.89	3.54	91
6	24–26	4.20	3.82	91
7	26–28	4.04	3.72	92
8	28–30	7.24	5.94	82
9	30–32	11.47	6.65	58
10	32–38	13.12	6.19	47

Lithology of Profile

Sample No.

1-3 Fine sandy loam, brownish gray.
4-7 Fine grained sand, light buff.
8-10 Silty clay, buff to brown.

TABLE 5.—*Distribution of carbonates in the dolomitized silts*

Three-quarters of a mile west of southeast corner of sec. 34, T. 144 N., R. 46 W., in Norman County, Minn.

Sample number	Depth in inches	Distribution of carbonates (Per cent)		
		Total carbonates	Total dolomite	Proportion of carbonates in form of dolomite
133	0– 6	1.95	1.38	70.7
134	6–12	2.92	1.84	63.0
135	12–18	3.68	2.21	60.0
136	18–24	23.72	16.01	67.5
137	24–26	23.88	17.53	73.4
138	26–28	21.84	18.06	82.7
139	28–30	21.74	19.00	87.4
140	30–32	23.61	21.53	91.2
141	32–34	22.26	20.15	90.5
142	34–36	22.79	20.62	90.5
143	36–38	21.90	20.37	93.0
144	38–40	21.07	19.41	92.1
145	40–42	22.26	20.15	90.5
146	42–44	22.44	20.42	91.0
147	44–46	27.51	21.87	79.5
148	46–48	31.69	24.84	78.4
149	48–54	31.44	21.47	68.3

Lithology of Profile 21

Sample No.

133-135 Dark silty clay.
136-138 Gray silty loam.
139-146 Very fine sand, buff with brown spots and streaks.
147-149 Silty clay, grayish green.

PETROGRAPHY OF SILTS

The sediments are of very uniform texture. A screen analysis of a composite sample from Profile No. 4 is plotted graphically in Figure 6. More than 50 per cent of the silt is finer than $\frac{1}{16}$ mm. in diameter.

The fraction finer than $\frac{1}{16}$ mm. was analyzed by the Krumbein pipette method with the following results:

Grades (mm.)	Per cent
1/16 –1/32	40
1/32 –1/64	32
1/64 –1/128	18
1/128 –1/256	6
1/256 –1/1000	3
1/1000–1/2000	1

In view of the fact that no more than 1 per cent of the sample was finer than $\frac{1}{1000}$ mm. in diameter, the material below that size was not fractioned any further.

The predominant minerals are quartz, feldspars, and carbonates (Pl. 1). Heavy accessory minerals make up less than 1 per cent of the sediments. The frequency distribution of the heavy minerals is shown in Table 6. The preponderance of hornblende suggests a considerable contribution from the pre-Cambrian crystalline rocks of the southwestern part of the Canadian shield.

TABLE 6.—*Petrographic analysis of Samples*
From Profile 4, SE. Cor. sec. 18, T. 140 N., R. 46W.

Sample Number	Depth in Inches	Size of Grain (in mm)		Wt. % Heavies	Distribution of Heavy Minerals												
		+1/16	-1/16		Apatite	Biotite	Epidote	Earthy Agg.	Garnet	Hornblende	Opaques	Pyroxene	Tourmaline	Tremolite	Zircon	Unknown	Misc
17	10	48	52	25	4	1	7	6	2	56	1	5	9	2	5	6	Ph. X
18	16	58	42	57	2	0	6	5	4	58	2	3	4	3	7	5	Sp. X Zo. X
19	22	54	46	62	5	3	5	5	4	55	3	10	1	2	4	2	R. X Zo. X
20	26	30	70	49	4	2	4	4	3	66	3	4	1	3	3	3	R. X
21	28	18	82	21	3	3	9	4	2	56	4	4	3	3	4	9	Ph. 2
22	30	22	78	47	4	2	6	3	3	63	2	4	1	2	4	3	Ph. 3
23	32	38	62	77	3	3	7	4	3	63	3	3	2	2	2	4	Ph. 2
24	34	28	72	1.07	2	3	7	2	5	62	6	3	1	2	1	4	Mon. X
25	36	26	74	1.29	5	2	8	2	4	59	4	3	3	1	3	3	Ch. 2
26	38	26	74	1.55	4	3	9	5	5	56	4	4	2	3	2	2	Za. X Mon. X
27	40	36	64	1.69	4	1	5	4	8	59	4	5	3	1	3	3	Ch. X
28	48	28	72	79	3	1	6	4	5	61	4	4	4	1	2	4	Ch. 1
29	60	45	55	35	3	1	8	3	2	64	2	3	1	1	3	3	Ph. X Ch. 1
30	72	40	60	1.03	7	1	9	3	3	61	3	3	1	1	3	4	Ph. X
31	96	30	70	.03	1	0	9	5	5	48	7	2	1	1	4	6	? 9
32	120	31	69	.03	1	1	6	5	9	26	11	2			3	4	Ch. 3 Ph. 28

Ph = Phosphatic Concretions
Sp = Sphene
Zo = Zoisite
R = Rutile
Mon = Monazite
Ch = Chlorite

MODE OF OCCURRENCE

Dolomitization has taken place where the water table is near the surface or where the natural drainage is poor, but not typically swampy. There is some enrichment also under the gentle slopes around swampy areas, but no dolomitization was observed where there is good natural drainage. In many localities the dolomite is associated with off-shore sands that dovetail with thin layers of silts and clays. In these sedi-

ments the water is under hydrostatic pressure, and thus the water table is near the surface.

The dolomitic bodies are not simply a product of negative enrichment formed by the leaching of calcium carbonate from the silts by vadose

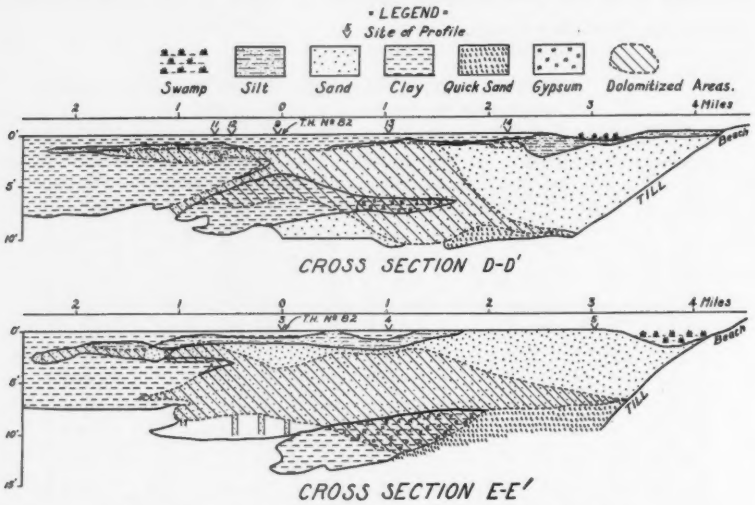

FIGURE 4.—*Cross sections along D-D' and E-E' of Figure 3*

circulation. In several places the dolomite occurs immediately below highly calcareous sands and it was found also under impervious clays.

The boundaries of the enriched zones do not follow the stratification of the silts but penetrate the bedding at various angles. The upper surface of the dolomite is generally roughly parallel with the surface of the ground and in most places begins from 12 to 30 inches below the present surface. The areas are very irregular in outline and in total thickness, and the thickness varies greatly within short distances (Figs. 4 and 5).

Horizontally the enriched areas vary from a few square yards to several acres. One exceptionally large area covering about 36 square miles occurs near the center of Clay County (Fig. 3). In this deposit the dolomite zone is from 7 to 10 feet thick. The total area that has been dolomitized is only a small fraction of the entire lake basin. The areas are scattered widely, and most of them are small. The relationships shown in Tables 1 to 5, and Figures 4 and 5 are typical. Many other profiles were studied.

The upper and lower limits of typical dolomitized zones are shown in Tables 4 and 5 which represent profiles that were sampled at short verti-

cal intervals. In Table 5 the zone with 88 per cent or more of the carbonate present in the form of dolomite begins at 29 inches from the surface and extends to a depth of 44 inches. Below that depth the percentage of dolomite decreases gradually.

At this location the dolomite enrichment is in a light, buff-colored, fine sand that overlies silty lacustrine clay. This area is small horizontally and has distinct boundaries. The surface soil contains a visible amount of soluble sulphates—mainly magnesium sulphate—that has been concentrated and precipitated by evaporation.

Profile No. 1 (Table 4) represents the vertical relationships under the center of a shallow topographic basin that covers an area approximately 70 feet by 350 feet. The enriched dolomitized zone begins 20 inches below the surface and is 8 inches thick. The overlying subsoil and the underlying clay effervesced violently with cold dilute hydrochloric acid. Samples obtained by borings and tested in the field showed that the dolomitic enrichment occurs under all of the depression and wedges out toward the margins of the shallow basin.

Table 4 shows that even though the percentage of total carbonates in the silts is small (0.86 to 13.12 per cent) there has nevertheless been concentration of dolomite under the poorly drained area.

ORIGIN OF THE DOLOMITE

The mode of occurrence of the dolomitic zones in the sediments precludes any possibility of mechanical concentration during sedimentation. Most of the silts are finely laminated, and some of them are distinctly varved, implying slow settling from quiet water. Furthermore, as stated, the dolomitic zones are not stratigraphic units. The possibility of leaching and reprecipitation in the sediments by the fresh melt waters of Lake Agassiz likewise seems quite remote.

In view of the fact that the ground waters in the silts are highly saline (Table 7), it seems more reasonable to assume that processes similar to those that are thought to take place in unconsolidated marine sediments on the floor of the sea may now be taking place in the lacustrine sediments of Lake Agassiz. The dolomitic zones in the silts may have been formed by the reaction of magnesium salts on the primary calcium carbonate (Parsons, 1922) that is abundant as detrital grains in the lake sediments. The presence of numerous nodules of gypsum crystals in the silts and clays below the dolomitic zones is evidence in favor of such an interpretation (Clarke, 1924, p. 586).

Table 7 gives partial analysis of ground waters and river waters of the Lake Agassiz region (Allison, 1932). One is at once impressed by the high proportion of salts of magnesium. Both $MgCl_2$ and $MgSO_4$ are

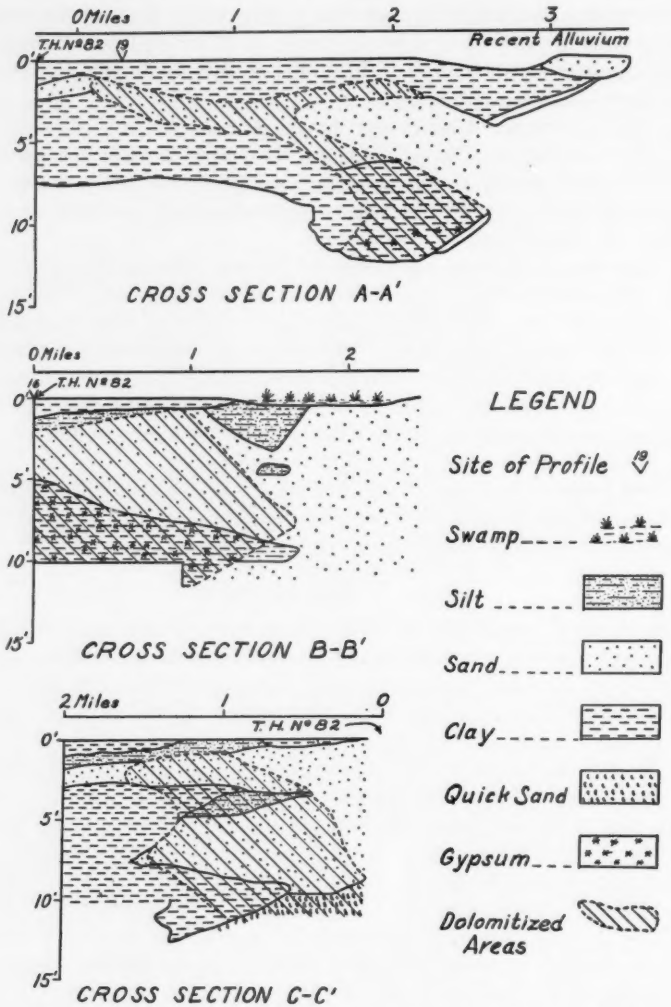

FIGURE 5.—*Cross sections along A-A', B-B', and C-C' of Figure 3*

present and undoubtedly both played a part in forming the dolomite. However, since NaCl is also present in the waters that saturate the sediment, it may tend to repress the ionization of the $MgCl_2$ and thereby lower the efficiency of the magnesium in $MgCl_2$ in replacing the calcium

in calcite (Van Tuyl, 1914). The extent to which such influence would retard dolomitization, however, is questionable.

TABLE 7.—*Analyses of Waters from the Lake Agassiz region*
Compiled from Allison (1932)

	Parts per million							
	1	2	3	4	5	6	7	8
Depth in feet	15	80	50	10	16	22	.....	.....
SiO_2	24	14	20	13	31	17	nd.	nd.
Fe	1—	1	5	1—	1	tr.	nd.	nd.
Al	2	11	1	2	22	3	nd.	nd.
Ca	183	500	315	99	169	128	65	75
Mg	156	700	223	30	90	144	36	25
Na	63	278	151	50	73	60	20	678
K	15	19	39	56	13	30		85
HCO_3	742	581	598	259	234	737	289	348
SO_4	281	3581	1452	169	599	256	80	457
Cl	142	245	11	25	3	75	6	792

1. Dug Well NW. ¼ sec. 10, T. 139 N., R. 48W.
2. Bore Well N. ½ sec., T. 139 N., R. 48W.
3. Well NW. ¼ sec. 34, T. 135 N., R. 47W.
4. Driven Well in Gary T. 145 N., R. 44W.
5. Bored Well NW. ¼ sec. 17, T. 143 N., R. 47W.
6. Dug Well in village of St. Vincent.
7. Average composition of waters from streams of Red River Basin.
8. Average composition of waters from Cretaceous deposits.

The data relating to the relative solubilities of the carbonates of calcium and magnesium are not wholly in accord. It has been found that calcareous sands on the shores of the Bahamas exposed to the action of sea water had their magnesium carbonate content leached more actively than the calcium carbonate. The work of Skeats (1905) indicates, however, that in fresh carbonated waters calcium carbonate may be leached away or replaced in part by magnesium carbonate. Leather and Sen (1914) state that calcium carbonate becomes practically insoluble in the presence of an excess of magnesium bicarbonate, and a mingling of solutions of calcium carbonate and of magnesium carbonate results in the precipitation of most of the former. They report further that if a

"mixture of calcium and magnesium carbonates is subjected to the action of carbonic acid and water, the calcium carbonate is largely or wholly prevented from dissolving."

Their studies showed that dolomite dissolves in carbonated waters in the form of the double carbonates.

The analyses of silt waters listed in Table 7 are typical of the water in shallow wells in the region. In the poorly drained, shallow topographic basins the salinity of surface waters is greater, due to concentration by evaporation. Locally the concentration may have reached a point where

magnesium carbonate was dissolved more rapidly than calcium carbonate near the surface. In other locations the calcite was leached more readily than the primary dolomite.

In the replacement of calcite by dolomite, the fineness of grain and the porosity also are important. Other things being equal, the more

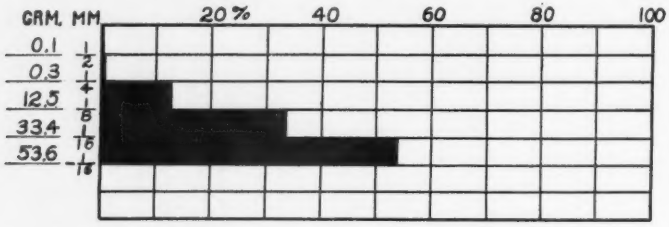

FIGURE 6.—*Graphic representation of screen analysis of a composite sample of silt*
From Profile Number 4.

porous the sediment and the finer the grains, the more susceptible it is to dolomitization. The lacustrine silts of Lake Agassiz are very well sorted and contain very little clay that would tend to clog the pore spaces. This assures greater permeability and consequently more rapid solution. Furthermore, large areas of the silt are overlain by a thick bed of peat or by a heavy layer of humus. The CO_2 generated by the decay of this organic matter aids in taking $CaCO_3$ into solution. Thus anything which tends to render the calcite more soluble will hasten replacement.

The transformation of calcareous sediments into dolomitic sediments is at best a slow process. In the lacustrine silts of Lake Agassiz the replacement was accomplished under very stagnant conditions. In such an environment the transfer of fresh magnesium salts to the points of reaction must have been accomplished by a process of diffusion which operates very slowly. The actual length of time during which dolomitization could have taken place may be estimated from the glacial history of the region. Obviously very little if any replacement of calcite by dolomite took place during the time that fresh water derived from the melting glacier occupied the lake basin. The present salinity represents postglacial concentration.

Leverett estimates the beginning of the south end of Lake Agassiz as taking place 18,000 years ago. If that time interval is of the proper magnitude, Lake Agassiz drained northward during the past 10,000 to 15,000 years. In that length of time the diagenetic changes outlined in the present paper have taken place.

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GEOLOGY OF ULUPAU HEAD, OAHU

BY CHESTER K. WENTWORTH AND J. EDWARD HOFFMEISTER

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ABSTRACT

Ulupau Head is a secondary tuff crater forming the eastern salient of Mokapu Peninsula, northeast coast of Oahu, Hawaii. Subsequent to its building from an offshore, submarine vent, the crater was deeply eroded by wave action on the north and east sides, the rim being largely destroyed on the east, and the sea entering the bowl. Later, at a stand of the sea at least 45 feet above the present stand, a bed of oyster-bearing limestone and an overlying calcareous beach sandstone were deposited on steeply sloping rock surfaces inside the crater.

Unconformable on these beds is a thick series of alluvial, tuffaceous silts, and in one sector a remarkable giant breccia, carrying 30-foot blocks of tuff and probably due to an enormous rock fall from an oversteepened marine cliff. At this time the

sea stood at least as low as at present. Later reef formations and accompanying beach conglomerates indicate two more stands of the sea, the first at 25 feet and the second at about 12 feet above the present.

Incorporated in the reef limestone of the northwest shore are great blocks of tuff let down by marine erosion from the rock-fall breccia and forming a remarkable reef breccia. This in turn, subject to attack by water-level weathering on the tuff blocks and solution benching on the reef limestone matrix, furnishes a striking example of the contrasted features of marine benches developed by processes peculiar to the two types of rock.

INTRODUCTION

LOCATION

Ulupau Head, an eroded tuff cone, is one of two projecting points of Mokapu Peninsula, which forms a prominent salient on the northeast coast of the Island of Oahu, Hawaii (Fig. 1). This salient extends northeast about 4 miles beyond the general windward shore line and is about midway of the southeastern and most eroded half of the windward slope of the Koolau Range. In contrast to the northwestern half, on which there is a fair preservation of the original form of the dome, the southeastern half has been deeply dissected to expose remnants of the deeper volcanic structure.

PURPOSE AND SCOPE OF STUDY

This paper represents a merging of two distinct studies. Hoffmeister, during a recent sojourn in Honolulu, has supplemented extensive investigations of corals and Pacific coral islands by comparative studies of reefs and reef limestones on Oahu. Wentworth has for several years given much holiday time to an intensive study of marine bench-forming processes on limestone and other rocky shores. A field review and discussion of the remarkable geomorphic and stratigraphic features of the Ulupau area suggested a further clinical study of the area, combining analyses of processes with as much historical interpretation as might prove practicable.

HISTORY AND ACKNOWLEDGMENTS

Ulupau Head has long been known as belonging to the series of secondary tuff craters of southeastern Oahu, of which Diamond Head is the most familiar example. Dana recognized Ulupau Head as a tuff crater similar to Diamond Head and mentioned also three other neighboring hills on Mokapu Peninsula which can readily be identified as Puu Hawaiiloa, Pyramid Rock, and Pali Kilo. He regarded these as craters or vents from which lava had been erupted (Dana, 1891, p. 299-300). Hitchcock was evidently aware of the character of Ulupau Head, since in listing tuff craters "probably not all active at the same time," he includes the "Kaneohe group" (Hitchcock, 1911, p. 43). In a report on the pyroclastic craters of southeastern Oahu, Wentworth (1926, p. 89-90) devoted five pages to the Ulupau group, describing the general

form, structure, and stratigraphy of the seaward part of Mokapu Peninsula. Stearns, in 1935, published additional details of observation and of correlation with stands of the sea worked out by him in connection with his areal and ground-water mapping of Oahu (Stearns and Vaksvik,

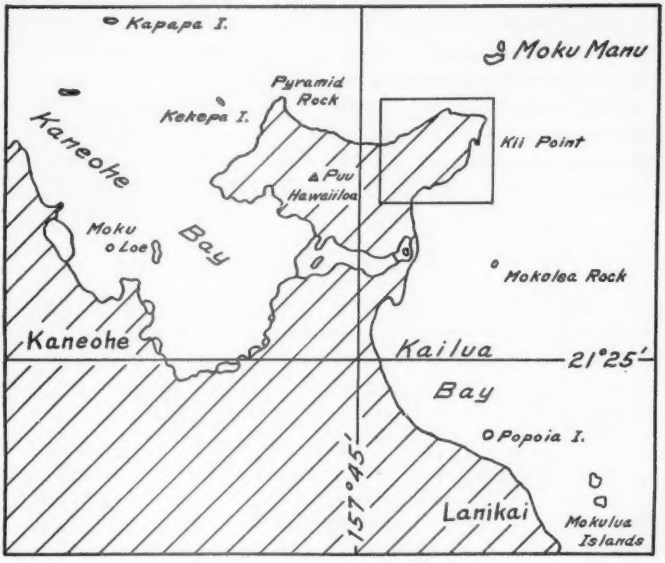

FIGURE 1.—*Index map*

Showing Mokapu Peninsula in relation to adjacent parts of Oahu. Outlines of geologic map (Fig. 2) are indicated.

1935, p. 99-101, 120-123). He also mentions the Ulupau Head area in a paper on shore benches (Stearns, 1935, p. 472). Interpretations by these earlier writers will be alluded to in the present paper.

The present study has been assisted in minor ways through the courtesy of the University of Hawaii, the University of Rochester, the Bishop Museum, and the Honolulu Board of Water Supply. Access to parts of the Mokapu and Ulupau areas was facilitated by special permission cordially granted by H. K. L. Castle, Manager of Kaneohe Ranch Company, and Delegate S. W. King.

PHYSIOGRAPHY

GENERAL TOPOGRAPHY

At the point where the geologically recent part of Mokapu Peninsula joins the older mass of the island, the peninsula is slightly over a mile

wide. This portion is cut obliquely by a large, shallow inlet, known as Nuupia Pond. This inlet is cut off from the ocean on the eastern side by a beach ridge about 10 feet above sea level and about 300 feet wide. On the western side it is separated from Kaneohe Bay only by an arti-

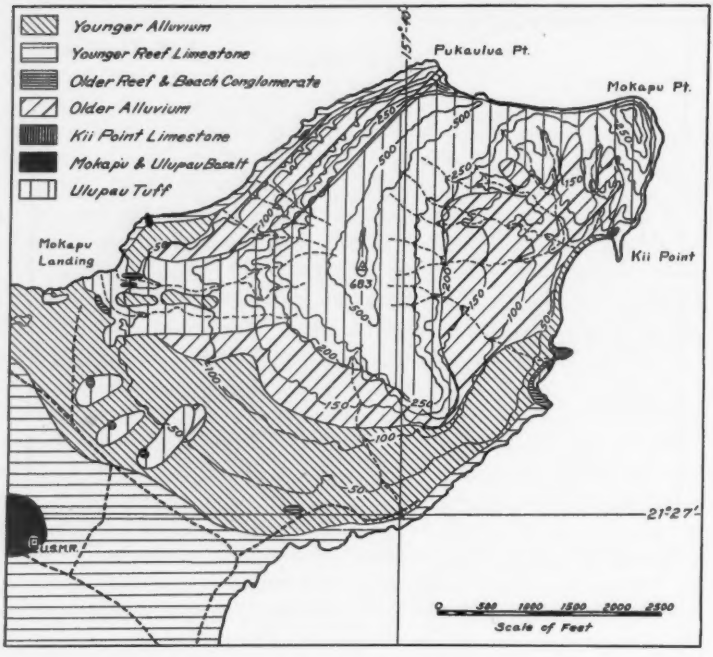

FIGURE 2.—*Geologic map of Ulupau Head*

ficial dike. Branches of the inlet or swampy areas extend to the north and to the southeast, and all represent parts where the peninsula has not been built above present sea level.

The peninsula widens at the northern end to a total of 3½ miles between the northeastern promontory of Ulupau Head, Mokapu Point, and a low tip of land west of Puu Hawaiiloa. Three types of highland rise above the general lowland of the peninsula: (1) The tuff crater of Ulupau Head; (2) the cinder and lava cone of Puu Hawaiiloa with its associated lava flow masses; and (3) the Mokapu shore cones which lie inland from the north-facing shore west of Ulupau Head. Except for these masses the Mokapu plain is nowhere over 15 feet above sea level.

Ulupau Head stands on a subcircular base, once over a mile in diam-

eter, from which parts have been cut away by the sea on the east, north, and northwest sides. The highest part consists of a crescentic remnant representing approximately the western half of the original crater ring (Fig. 2; Pl. 1, fig. 1). This remnant reaches an elevation of 683 feet at a point lying about S. 80° W. of the probable position of the vent. A smaller part of the original northeastern rim which forms Mokapu Point and Kii Point reaches an elevation of 328 feet. Inner and outer slopes of the higher parts of the tuff cone approximate 35 degrees and like many of the Oahu tuff cones have been stripped and somewhat furrowed, but not deeply dissected. Farther down, the slope merges into the lesser declivity of a mantle of alluvial fans, separated by the channels of a few gulches most of which reach maximum depths of less than 30 feet (Fig. 2).

SHORE BEACHES AND BENCHES

If one approaches Ulupau Head from the south, the eastern shore of Mokapu Peninsula is seen to consist of a sand beach, continuous to a point about three-quarters of a mile south of the highest point of the western rim. Here is the beginning of a limestone shore on which is developed a typical solution bench,[1] bordered inland in most places by an overhanging nip and visor cliff 4 to 6 feet high, behind which is a pitted zone on the emerged reef surface. This type of shore swings eastward and northward around the southern end of the main remnant of Ulupau crater toward Kii Point. At about 2000 feet from Kii Point the reef-rock-solution-bench shore gives way to shore cut into the tuff of Ulupau Head and associated rocks.

At Kii Point is a large area of water-leveled bench [2] cut on a remnant of tuff which swings southeastward from the point and joins the line of remnants mentioned; this line probably marks the structural axis of the crater rim. From Kii to Mokapu Point the shore is a vertical or overhanging cliff, 50 to 200 feet high and trending northward. At Mokapu Point the shore turns westward as a similar cliff which reaches 500 feet in height at one point and is the steepest cliff anywhere on the Oahu coast. The base of this cliff generally runs off into deep water, with only a slight curve flaring outward at water level. This stretch of coast is quite inaccessible from the land side, but its chief features can be clearly seen from a boat. The western end of the high cliff is Pukaulua Point.

The northwest coast of Ulupau Head is chiefly composed of limestone with a solution-bench shore cut in rock whose pitted surface rises to about

[1] The term solution bench has been used by Wentworth (1939) to identify the low bench produced on limestone coasts by the solvent action of fresh water.

[2] Water-leveling is the term used by Wentworth (1938) to describe certain benches on the coasts of Oahu, Hawaii, produced by standing sea water caught in pits and irregular depressions in volcanic rocks.

10 or 12 feet in elevation. Just north of Mokapu Landing coral reef and detrital limestones predominate. In the vicinity of Pukaulua Point one notices occasional blocks of tuff in the reef limestone of the pitted zone and also in the same limestone of the solution bench.

As the abundance of such blocks increases to form a conspicuous part of the mass, the peculiarities of response of the tuff to weathering and the bench-forming processes become apparent. Both in the pitted zone and in the solution bench, the tops of the tuff blocks are flat and level, despite the fact that the bedded structure of the tuff is oriented quite at random. Everywhere, the leveled surfaces bevel the bedding. On the solution bench, the top surfaces are approximately at the same level—some few inches but less than a foot above the general surface of the limestone (Pl. 1, fig. 3). In the pitted zone, on the contrary, the leveled surfaces of the tuff blocks stand at elevations varying from a few inches to 3 or 4 feet below the level of the limestone surface (Pl. 1, fig. 2). The leveled tops of blocks, then, are level bottoms of pits in the reef mass which correspond in outline to the blocks themselves. The marginal walls of the flat-bottomed pits are almost wholly composed of limestone, though in places a small remnant mass of tuff may extend upward in the face of the vertical wall on one side or corner. Similarly, in a few places the tuff of the block may extend slightly under the limestone of the wall, showing minor discrepancies between the areal outline of the tuff block and of the pit itself, both considered at the level of the pit floor (Pl. 1, fig. 3).

In nearly all instances the pit displays an outlet at the level of the floor, either a gap or a roofed orifice, through limestone, into a lower pit or channel. To one familiar with the features of water-leveled benches, the observed relationships strongly suggest that the pits have been cut by water-level weathering, operating on the tuff of the blocks, just as it does on the tuff in place on various coasts (Wentworth, 1938, p. 6-32). The agreement between pattern of blocks and of pits leaves no escape from the view that, under the controlling conditions, the water-leveling of the tuff blocks has been more rapid than the lowering and destruction of the surrounding reef limestone by solution pitting, owing perhaps to the meager amounts of fresh water available here.

On the solution bench, interrupted by the remarkable, flat-topped tuff blocks, one can only reach the opposite conclusion that here the reduction of the solution bench on the matrix limestone must still be in progress at a slow rate, while the water-leveling of the tuff blocks, owing to complete lack of surrounding rims to retain the water, must be relatively inoperative (Pl. 1, figs. 2 and 3). There is no reason to suppose that the tuff blocks of the solution bench zone originally stood lower, generally, than those of the pitted zone. It seems reasonable to suppose that, as the lime-

FIGURE 1. VIEW OF ULUPAU CRATER FROM THE SOUTH

FIGURE 2. SOLUTION BENCH (MIDDLE GROUND) AND PITTED REEF SURFACE (FOREGROUND)
Water-leveled tuff blocks stand in slight relief on solution bench and depressed in reef zone.

FIGURE 3. EDGE OF PITTED REEF ZONE (FOREGROUND AND RIGHT)
Along solution bench cut on reef limestone breccia.

ULUPAU HEAD AND SHORE BENCHES

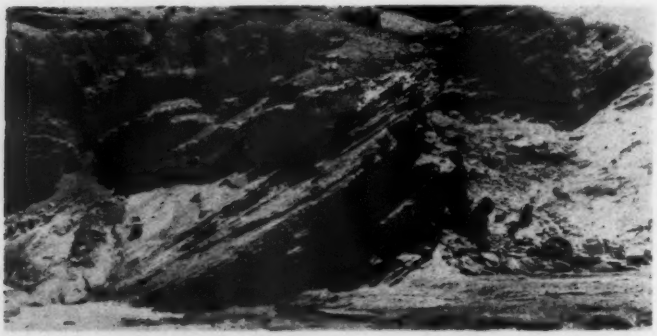

FIGURE 1. KII POINT LIMESTONE (WEDGE OF LIGHTER BEDS)
Lying on dipping beds of tuff (center and right) and overlapped by older alluvium.

FIGURE 2. OLDER ALLUVIUM
Showing calcareous stem and root casts in cliff near Kii Point.

FIGURE 3. ROCK-FALL BRECCIA
In cliff northwest side of Ulupau Head.

FEATURES OF SEA CLIFFS AROUND ULUPAU CRATER

stone of the reef matrix was reduced to a level approaching the solution bench equilibrium, its efficacy as a rim for the retention of water on the tops of the tuff blocks would be nullified and the leveled tops of the tuff blocks would tend to be left at subequal elevations as they are now found. In a recent paper on shore benches of Oahu reference is made to

> "the northwest side of Ulupau Head [where] the lower bench truncates a talus breccia consisting of blocks of tuff cemented in a matrix of fine tuffaceous silt . . . this lime-leached matrix stands in relief as pinnacles several inches high, whereas the limey tuff blocks have been dissolved out; hence, solution must be more rapid on this particular bench than erosion." (Stearns, 1935, p. 1472.)

The present writers were not able to find in the bench any breccia with a lime-leached, tuffaceous matrix; the breccia exposed in the bench, as already stated, is a reef breccia, composed of tuff blocks in a limestone matrix, the talus breccia occurring only in the high cliff behind the bench. The relief, both on the upper platform, which is the top of the reef itself, and the lower solution bench cut in the reef breccia, is due to differential effects of the water-leveling and solution-benching processes. The writers agree with Stearns that these are more rapid here than abrasion or quarrying ("erosion").

AREAL AND STRUCTURAL GEOLOGY

GENERAL STATEMENT

Interest in the Ulupau Head area as expressed in the present paper centers around the stratigraphic succession of formations and in their relations to the chronology of sea-level stands. As a relatively small mass of only moderately resistant rock, standing exposed to the successive attack of seas at several levels, Ulupau Head, like other tuff cones, is peculiarly well situated to record a succession of geologic conditions.

ULUPAU TUFF

The Ulupau tuff has already been described in some detail, being similar to the other secondary palagonitic tuffs of southeastern Oahu (Wentworth, 1926, p. 85-90; Stearns and Vaksvik, 1935, p. 121-122). It tends toward gray rather than brown tints, owing to the large amount of reef detritus and secondarily deposited lime it contains. Petrographically, it belongs to the Honolulu series and contains both nephelite and mellilite. The structure of the crater is typical of other such cones which have already been described. It is noteworthy that the east-facing, inside slope of the main mass shows a marked exposure of the edges of strata which belong to the outward-dipping structure of the rim. Thus it is suggested that a considerable portion of insloping beds of the inner part of the west rim has been removed by erosion. This fact may help to explain the large amounts of alluvial material now deposited inside the rim.

MOKAPU BASALT AND ULUPAU DIKE

Near the middle of the reef rock and solution bench lobe of the shore which lies northeast of Mokapu Landing, a small mass of porphyritic basalt of the Honolulu series protrudes through the reef rock of the solution bench. Stearns (1935, p. 101), who did not record this exposure in the bench, has reported two small masses of basalt lying about 4000 feet and 9500 feet roughly southward of this point and has given to these the name of Mokapu Basalt. The latter are vesicular, somewhat flow-banded, and clearly of extrusive character. On the contrary, the basalt of the outcrop on the solution bench is dense. It does not have the regular columnar jointing of some dikes, but rather the irregular jointing into blocks which suggests derivation from a dike vent very close to its present position.

KII POINT LIMESTONE

At Kii Point the indipping tuff beds of the Ulupau rim dip southward at an angle of about 26 degrees. Resting directly on this slope is a wedge-shaped mass of limestone made chiefly of oyster and barnacle shells. At the level of the sea the limestone is about 20 feet thick and then pinches to nothing at a height of 45 feet. From their position it is clear that the limestone beds were deposited on a rather steeply dipping, rocky bottom which at the time was swept clean of debris. The lower 6 feet consists of irregularly imbricated oyster shells encrusted one on another, with a large number of barnacles also included in the mass. In the 2 to 4 feet next above, the oysters become less numerous, their place being taken by the barnacles. The formation then grades to a calcareous sandstone composed of beach detritus (Pl. 2, fig. 1).

At a point about half a mile to the south the Kii Point limestone also outcrops in a similar relation to the crater rim, resting as it does on the indipping (here northward) tuff beds. Here the total thickness of this calcareous series is about 30 feet. The lower 2 to 3 feet is composed of oyster shells of the same species, above which is 25 to 28 feet of a greenish-yellow, calcareous detrital beach sand, in most places poorly cemented. This sand is made of Foraminifera, small fragments of many other types of calcareous organisms, tuff grains, and a few large shells in excellent state of preservation.

The commonest of the larger shells are the conical gastropod *Helcioniscus exaratus melanostoma* Pilsbury, the pelecypods *Chama iostoma* Conrad, *Ostrea hanleyana* Sowerby, and *Antigona reticulata* (Linnaeus), and a few massive corals.[3]

A bone found partly imbedded in this sand formation and clearly of

[3] The writers are greatly indebted to J. M. Ostergaard of the University of Hawaii for the identification of the Mollusca mentioned in this paper.

contemporaneous deposition has been identified by Dr. Remington Kellogg of the United States National Museum as the first sternal rib, right side, of a blackfish, *Globicephala,* which is a type of small whale. From the single bone it was not possible to determine whether or not the species is the same as the living Pacific blackfish, *Globicephala scammonii.*

At one point in this vicinity a basal tuff conglomerate is firmly cemented against the eroded surface of the tuff. It is not clear from the relation whether this is basal to the Kii Point formation or if it is a part of the younger transgressive conglomerate described below.

The Kii Point limestone as defined here includes the oyster and barnacle beds, which rest on the tuff, and the overlying detrital calcareous beach sandstone. Unconformably above the beach sand member of the limestone is a thick deposit of tuffaceous sediments which will be described as the older alluvium.

Stearns (Stearns and Vaksvik, 1935, p. 121-123), who has described this locality, states that the oyster and barnacle limestone is "overlain by several beds of fine, greenish-yellow silt which lie conformably on it". Since the beach sand and silt of the Kii Point formation and the overlying beds of the older alluvium are both greenish-yellow, it is logical to assume that Stearns considered them parts of the same formation and did not recognize the unconformity between them. He also states that,

> "the nature of the contact of these beds (yellow-green silt) with the adjacent nearly horizontal sediments (main, older, tuffaceous, secondary beds in Ulupau Crater) is indefinite, but the uppermost of the steeply dipping beds seems to merge into the horizontal beds."

The intent of this statement is confirmed by the section contained in Figure 10 of Stearns' report, in which the whole series, from the dipping oyster beds to the uppermost of the main crater-filling beds (labeled "tuffaceous marine sediments, possibly alluvium at top") is shown as conformable on the south side but with an angular break at the top of the limestone beds at the north end. As will appear in the next section, the present writers do not fully accept this interpretation.

The Kii Point limestone also can be recognized on the opposite or northwest side of Ulupau Head. Here along the base of the cliff at the back of a 12-foot limestone platform and underlying the older alluvium there are two exposures of the beach sand containing the characteristic shells. At first sight the position of the sand suggests the possibility of its being contemporaneous with the 12-foot reef flat. However, its position directly beneath the older alluvium of this side and the fact that its organic composition is distinct from the beach deposit found at other places along this coast, which is contemporaneous with the 12-foot reef flat, demonstrates its separation from this younger formation.

OLDER ALLUVIUM AND ROCK-FALL BRECCIA

Above the detrital upper beds of the Kii Point formation on both the east and the northwest coasts of Ulupau Head is a thick series of tuffaceous sediments which appear to be of subaerial origin. On the east side these beds are exposed in the cliff extending about 1200 feet southwestward from Kii Point and are composed of fine dust and silt produced by the disintegration of the Ulupau tuff. The structure is essentially massive, except as it is given the marked appearance of stratification by lines of angular fragments of tuff and of coarse lapilli from the tuff (Pl. 2, fig. 2). The bedding is parallel, and no cross bedding or lensy type of lamination was noted. As Stearns has stated, the material might easily be mistaken in weathered outcrop for primary tuff, except for the lack of bomb sags and the presence of tuff fragments as pebbles and the like. In some places this formation is essentially free from pebbles or coarse fragments and in general it is greenish yellow or olive brown.

The beds of this formation overlie the detrital calcareous phase of the Kii Point limestone on the north side of a narrow coastal point 1200 feet southwest of Kii Point. The whole bank is washed by sea spray, and the contact between the two formations is not conspicuous. By digging it is readily shown that the soft, greenish-yellow, tuffaceous beds are separated from the lighter-colored, somewhat cemented, calcareous beds underneath by an extremely irregular line of contact. This contact appears to be an intricately pitted, in places overhanging, upper surface of the limestone with a relief of ruggedness of several feet. The silty beds were evidently deposited on such a surface producing an interlocking, jig-saw type of contact in vertical section which, because of the overwash of detritus from the upper beds, must be followed in detail by digging to verify the form suggested by irregular blotches of the lighter limestone.

No evidence has been found to indicate that these silt beds are marine, and there are several reasons for considering them to be of subaerial deposition. Close search from top to bottom showed no marine fossils. At many places throughout the body of the formation, but especially upward from 40 feet above sea level, there appear numerous, vertical, branching, secondary, calcareous casts which strongly indicate the growth of plants and shrubs. The casts were produced by circulating waters which passed along the stems and roots of the plants. Close study of these casts shows that they are identical in structure and distribution with root casts found in many other localities in Hawaii (Pl. 2, fig. 2). Some of these branching casts are as much as 3 inches in diameter. The rather uniform, wide strewing of coarser tuff fragments in lines and thin zones through the whole width of the formation as exposed in the sea

cliff and in the gullies which trench it inland is contrary to the writers' understanding of marine sedimentation. In marine beds such pebble accumulation, even if not rounded or sorted, would at least be emplaced transgressively with some evidence of fore-set bedding. When the beds are traced inland up the gulches to the tuff bedrock, there is no evidence of a break in type of stratification. These facts, together with the markedly unconformable contact described, enforce the conclusion that this whole series is not of marine but of fluviatile and colluvial origin. This older alluvial formation includes practically the whole sedimentary series, almost to the grass roots at the northern end of the crater-filling mass. However, southward in increasing thickness up to 10 or 15 feet in places, there is at the top a younger alluvium which can be separated from the older and which will be described.

On the northwest side of Ulupau Head fine-grained, tuffaceous alluvium of this older formation is exposed in the lower part of the cliff which rises from the back of a limestone shore platform. This formation nearly everywhere appears to pass down behind the limestone, but in two places its base rises high enough to reveal the underlying marine sand formation of the Kii Point limestone previously described. The finer alluvial beds are overlain by a thick mass of very coarse breccia or weakly cemented rubble, the elevation of the contact ranging from 20 to 50 feet along much of the bluff from Pukaulua Point toward Mokapu Landing. The finer alluvium on the northwest coast is very similar to the alluvium of the Kii Point coast and shows similar features which indicate a subaerial origin.

The coarse breccia of the upper part of the northwest Ulupau bluff is indeed a remarkable formation (Pl. 2, fig. 3). It is largely composed of angular blocks of Ulupau tuff 5 to 10 feet across, and larger blocks, ranging to a maximum of at least 30 feet in length and width, are common. They are arranged with the bedding of the tuff at all angles, and the matrix around the larger blocks is a rubble of smaller, angular blocks and finer, silty detritus from the tuff. The formation is sufficiently indurated to stand in a cliff 50 to 80 feet high at angles from 60 to 90 degrees, but it is by no means a strongly lithified mass. No tuff, in place, is seen along the coast northwest of Pukaulua Point until one approaches Mokapu Landing. Inland from the landing a line of exposures of tuff extends toward the high slope back of Pukaulua Point which suggests the upper edge of a cliff against which the breccia formation lies. The most plausible explanation of these relations is that the breccia was the result of a tremendous rock fall, or series of rock falls, from a cliff cut by wave action against the northwest slope of Ulupau Head. The massive character of the breccia suggests its formation en masse rather than bed by

bed, and it is also entirely likely that such a cliff was cut by waves and thus prepared for such sudden collapse. The most probable sequence of events is indicated in an accompanying serial diagram based on available exposures along the coast and the upper slopes of the crater (Fig. 3).

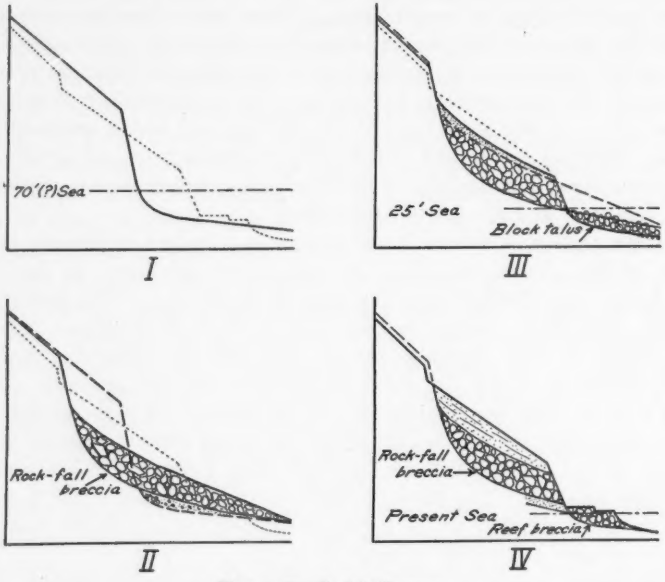

FIGURE 3.—*Serial diagram*

Illustrating sequence of events in evolution of existing profile and structure of northwest side of Ulupau Crater.

Undoubtedly, the breccia beds, as well as the underlying, finer phase of the older alluvium, formerly extended a considerable distance seaward of the present cliff.

Careful search along the northwest bluff also disclosed several small masses of reef limestone, in place and undisturbed, at elevations of 24 to 26 feet above sea level. These lie on eroded, downthrown masses of breccia and blocks from the breccia; there is thus no doubt as to the greater age of the breccia, masses of which were undermined by the advance of the sea in which the reef was later built.

OLDER REEF LIMESTONE AND ASSOCIATED BEACH CONGLOMERATE AND SANDSTONE

The conglomerate phase of the formation consists of well-rounded tuff cobbles up to 6 inches across, with a few boulders and blocks up to 2

or 3 feet in diameter, all fairly firmly cemented in a matrix of calcareous sand and secondary calcium carbonate. The matrix contains water-worn shells, chiefly coral fragments 1 to 3 inches in diameter. The conglomerate is transgressive on a previously eroded surface at elevations from below sea level to a maximum of 29 feet above sea level. Associated with it in places, particularly at its seaward exposures, are patches of coral reef rock grown in place. The largest of these is at the shore salient 1800 feet south of Kii Point where the reef rock is 12 feet thick and extends lengthwise along the beach for 75 feet. Of the corals large masses of laminar *Porites* and heads of *Pocillopora ligulata* Dana predominate.

In its upper parts the conglomerate bed is about 11 to 12 feet thick. The maximum elevation of the top of the conglomerate is 40 feet. At the inland and upper limit of the conglomerate it grades into calcareous beach sand, and this in turn in places grades into lenses of nearly pure olivine sand like that of parts of the modern beach. Near the northern end of the exposure of this formation the calcareous sandstone phase is overlapped by younger alluvium. In successive diggings to determine that the alluvium overlies the beach sandstone there was encountered under the alluvium a layer of several inches of coarse olivine sand, which in turn lies above the calcareous and better-cemented sand. That the olivine sand was a phase of the beach formation is quite clear. That it never appears in an exposure is due to its uncemented and easily eroded condition.

This beach conglomerate clearly overlies the beveled edges of both the oyster bed and detrital phases of the Kii Point formation and also the older alluvium on which it is less markedly but still clearly unconformable. Its contact with the alluvium is in places marked by a layer of pelecypods, doubtfully identified by Ostergaard as a *Chama,* and calcareous algae an inch to a foot in thickness. This layer represents the base of the conglomerate and rests directly on the beveled surface of the alluvium.

The beach conglomerate is also found on the coast southwest of Pukaulua Point. Owing to the vertical attitude of the cliff and the erosion to which it has been subjected, only very small amounts remain in protected situations, plastered against the alluvium and disturbed blocks of the rock-fall breccia. Remnants of conglomerate and other indications of this strand level are found at elevations of 25 to 35 feet inland from Mokapu Landing and at several places where the newer alluvium has been stripped off the southern and southwestern slope of Ulupau Head. At the time of the formation of this conglomerate the sea must have been about 25 feet above the present level—the Waimanalo stand of Stearns (Stearns and Vaksvik, 1935, p. 123). During this stand the older rocks

were beveled, and the terrace was formed on which the conglomerate was deposited.

YOUNGER REEF LIMESTONE AND REEF BRECCIA

At certain points around Ulupau Head, as described, there are coral reef masses with upper surfaces ranging to 25 feet above sea level. These were contemporaneous with the beach conglomerate and were formed during the Waimanalo stand of the sea. They are regarded as older reef patches grown in a position fringing Ulupau Head but not widely enough to join Ulupau Head to Oahu.

The younger and lower reef limestone referred to here forms a platform, ranging from 5 to 12 feet above sea level, which fringes the northwest Ulupau coast from Pukaulua Point to beyond Mokapu Landing and forms much of the lowland of Mokapu south of Ulupau Crater and Puu Hawaiiloa. While the extent and relations of this reef limestone in the latter area indicate its growth in a sea approximating 12 feet above the present level, and militate against interpreting it as formed in the 25-foot Waimanalo sea, the most directly conclusive evidence is found a short distance southwest of Pukaulua Point. Along this coast for about 4000 feet the landward margin of the reef limestone in the pitted zone back of the solution bench reaches a maximum elevation of about 12 feet. Along part of this distance the limestone abuts against a cliff cut in older formations. In one place there is a fairly distinct nip cut against a displaced mass of breccia at about 14 feet above sea level and about 10 feet below the level of near-by remnants of older reef limestone belonging to the Waimanalo stand. On the top of the reef limestone at about 12 to 13 feet above sea level in many places are small patches of cemented calcareous sand and gravel in which rounded algal pebbles are conspicuous, the whole indicating a stand of the sea at about 12 feet above the present level (Wentworth and Palmer, 1925, p. 521-544).

A part of the reef formation which fringes the northwestern Ulupau shore and rises to 12 feet is typical reef limestone. In a few places near the west end it contains a few fragments of tuff or gives way to small inliers or mantling patches of sandstone or conglomerate containing volcanic detritus. However, as Pukaulua Point is approached, the reef rock is increasingly laden with tuff blocks, some at least 20 and 30 feet in diameter, until near Pukaulua Point there are places where it appears that well over half the mass consists of tuff blocks of various sizes.

Two possible explanations for the emplacement of these blocks in such a reef formation are worthy of discussion. One of these would involve the direct rolling out of these large blocks onto a reef mass lower than

the present, or onto the sea floor at this point. Following such accumulation the reef mass grew around the blocks so as to cement them into the formation found today. The other postulate is that the landslide mass was placed mainly on land, overlying the finer alluvial formation, as seen in section, but extending some rods farther seaward than at present. Following such landsliding, the sea cut back into the land, forming a cliff approximating the position of the present bluff at the inner exposed margin of the reef, and at the same time leaving a large mass of the coarser blocks, not quickly or wholly removed by wave action, which were presently incorporated into the reef mass as in the first case.

Certain facts appear to make the second postulate more probable. The finer alluvium which underlies the landslide rubble appears to have extended some distance farther seaward. The horizon of the contact would not pass below sea level for some rods (Fig. 3). Therefore, any blocks rolled directly out on what was probably the grade of the older formation would have passed some distance seaward of the present reef breccia and would lie above its probable base. Blocks of tuff in the reef matrix are found practically at the base of the present bluff which truncates the landslide rubble, and their emplacement here seems definitely to require a letting down. We have no reason for supposing that there was an ancient steep cliff along this line, down which the blocks could have fallen. Moreover, the close spacing and rather large size of most of the blocks suggest a sorting of the materials in strong wave action. Such sorting could have been accomplished by waves working on a mass of landslide detritus thrown directly into the sea, but the perfection of the process and the rather notable uniformity of width of the reef breccia with the similar uniformity and parallelism of the cliff make the sorting by marine encroachment on an extension of the existing section seem more probable (Fig. 3).

YOUNGER ALLUVIUM

The distinction between this formation and the older alluvium is well shown along the top of the bluff about 2000 feet southwest of Kii Point. The older alluvium is somewhat indurated and carries more coarse material. The younger alluvium is finer, darker, and more uniform in texture. Furthermore, it lacks the distinctive root casts so common in the older alluvium. It becomes thicker around the southern end of Ulupau Crater and is probably the only alluvial formation on the lower ground southwest of the crater. This is believed to be due to the formation of the lower reef subsequent to the deposition of the older alluvium and to the lack of any widespread basement on which the older alluvium could accumulate outside the crater bowl to rise above present sea level in this area.

In the general vicinity of Mokapu Landing and inland from the reef platform containing the Ulupau dike mass, there are bank exposures of younger, darker, nonindurated alluvium of the same formation. It appears in general that the accumulation of younger alluvium has not been active on the slopes of Ulupau Crater which are at present truncated by steep cliffs, cut either in tuff or in older alluvium. Such slopes either are too steep to the cliff edge or are trenched by gulches to which all drainage goes; in either case conditions are produced which are not suitable to the formation of any great thickness of recent alluvium over the general surface.

GEOLOGIC HISTORY

Geologic history recorded in the Ulupau area commences with the eruption of the Ulupau tuff. Prior to that event the Koolau range of Oahu had been formed by a long sequence of volcanic events and had been extensively and deeply dissected.

The eruption of the Ulupau tuff and formation of Ulupau Crater took place from a vent which perforated calcareous reef formations, as shown by the rather large amount of accidental reef detritus contained in the tuff. There is good reason to suppose that all the palagonite tuff cones of southeastern Oahu were produced by phreato-magmatic explosions in which steam played a large part and therefore that they took place from vents either quite near or below sea level. The formation of such symmetrical craters was thought by Bishop (1901, p. 1-5) and Wentworth (1926, p. 54) to have occupied short periods, perhaps no more than a few hours, as in the case of some known historic eruptions of this sort. Stearns (Stearns and Vaksvik, 1935, p. 136) takes exception to some of the assumptions and mathematical procedure used by the latter but agrees "that the eruption (of Diamond Head) was short-lived". The general thesis appears still to stand and applies to Ulupau Crater equally with Diamond Head and other ring-shaped tuff craters.

No clear exposure of a transition from submarine to subaerial tuff beds is known to the writers either in the cliffs of Ulupau Head or in any other mass of tuff on Oahu. It is difficult to believe that such beds would not be fairly well distinguished by attitude and structure, as well as by erosional discordances immediately at the level of the contemporary sea. Lack of evidence in the structure of the tuff of a demarcation in the portion now above sea level leads the writers to hesitate in accepting allocation of the Ulupau eruption to the Laie 70-foot stand of the sea. It seems equally likely that the sea stood as low or lower than at present during the Ulupau eruption. On this basis, and with reference to the scale of sea-level positions developed by Stearns, the Ulupau eruption may

have taken place at any time shortly before the Laie stand, possibly during a negative stand not yet defined.

Stearns regards both Puu Hawaiiloa and the Moku Manu Islands as dating from the Kaena stand, and hence considerably older than Ulupau Head. As yet the authors have made no detailed study of these areas; at any rate it appears likely that all, including Ulupau Crater, were in existence prior to the Laie stand.

Age relations of the Ulupau (dike?) basalt and the Mokapu basalt of the two areas described by Stearns are difficult to determine. The latter are considered by him to be as old as the Kaena sea stand, because of their much eroded character. Another factor to be considered is the possible form of the ocean bottom before and after the Ulupau and Moku Manu eruptions. Placement of the Mokapu basalt in the area south of Ulupau Head and of the Ulupau basalt at the shore point west of Ulupau Head both seem more plausible after the formation of Ulupau than before. At least these call for a rock surface not very much deeper than the present surface of the peninsula. It is hardly possible to say whether eruptions from the Ulupau Crater vent or perhaps from the Moku Manu vent were concerned in building such a surface.

The next event clearly recorded in the Ulupau region is the rise of the sea probably to the Laie level—about 70 feet above the present. Ulupau Crater may already have been considerably eroded but was undoubtedly subjected to its most severe marine and fluvial erosion at this time. The original height of the east rim is not known; it seems most likely that it was first broken down to permit entrance of the sea into the crater bowl in Laie time. From the position of the Kii Point limestone it is evident that a fairly broad breach was made. Probably at this time scouring of the inside of the crater led to the rather marked subaerial stripping of the indipping beds from the inside of the west rim which is still in evidence. There was also strong erosion of the northwest side of the cone. Two bits of evidence suggest the extent of such erosion. Along the upper part of the northwest slope is a cliff line, above which tuff in place is exposed and below which the surface reveals only the rock-fall breccia. This is thought to be the upper edge of the sea cliff formed during Laie time.

The other point is that, along the seaward edge of the solution bench cut in reef breccia, whatever pavement underlies the reef breccia can hardly be less than 20 or 30 feet below present sea level to have allowed the placement of the large blocks on it. Since the fine-grained alluvium lies in two places on earlier sand of the Kii Point formation, and this in turn is at least a few feet in thickness, it is likely that the tuff pavement had been cut at least to present sea level in Laie time. The simplest

explanation of these relations is to suppose that, during the Laie period of active marine attack, a submarine platform and sea cliff were cut, the platform at least to present sea level at the present northwest coast, and rising to a nip at 70 feet at the foot of the cliff whose upper edge shows in the present slope (Fig. 3).

During this period of marine erosion or during a brief period of quiescence which may have followed, the Kii Point oyster beds were accumulated and the associated calcareous beach sands deposited. There may have been a brief halt at a somewhat lower level than the Laie stand at this time.

Following this, the writers believe, the sea withdrew to some level below the present strand line. Stearns (Stearns and Vaksvik, 1935, p. 123), on the basis of work in other parts of Oahu, believes that the sea fell to what he calls the Waipio stand which is about 60 feet below the present level. During this withdrawal there was much subaerial erosion of the Kii Point limestone and marked aggradation, on the irregular surface thus formed, of the lower slopes of the inside of the bowl and the northwest outer slope by the tuffaceous alluvium (older alluvium). Vegetation grew on this alluvium during its deposition and eventually gave rise to the root casts so common there. During this period also there took place the great rock fall which the writers believe to be responsible for the notable breccia of the northwest outer slope. This is thought to have been due to the oversteepening of the cliff in Laie time, but not to have taken place until after the beginning of alluvial aggradation. It was possibly induced by exceptional ground saturation or an unusually severe earth tremor.

The next event in the history of the area was the rise of the sea to the Waimanalo stand which was about 25 feet above the present level. During this time the beach conglomerate and associated coral reef patches were deposited unconformably on the surface prepared by wave cutting in the previously formed rocks. Although the beach sand associated with this conglomerate rises to a height of 40 feet, the writers concur with Stearns in correlating it with this 25-foot stand.

Following the Waimanalo level the sea fell to a stand which is approximately 12 feet above the present level. During this time was formed the younger reef limestone which is so extensive around Ulupau Head and which forms so much of the lowland of Mokapu Peninsula. The matrix of the huge tuff blocks near Pukaulua Point is chiefly a phase of this reef though some breccia matrix may have formed at the Waimanalo level. The younger reef is composed of corals grown in place at least 12 feet above the present sea and therefore could not have been formed during a lower stand. The presence of sea-cut nips behind the reef, such as the one described, shows that it could not have been formed in the

higher Waimanalo 25-foot sea. The platform lowers to about 5 feet in the lowland area surrounding the Mokapu Peninsula and joining the higher salient to the mainland. It is possible that the sea, during its withdrawal to the present stand, stopped for a time at the 5-foot level. However, it is believed that the greater portion of this lower deposit was formed during the 12-foot stand. Because of the length of time required to produce the deposits formed during the 12-foot level and because of its clear-cut character, it is here designated the Ulupau Stand.

The last formation to be deposited was the younger, recent alluvium which is now most conspicuous around the southern end of Ulupau Crater and which is readily distinguished from the lighter-colored, older alluvium.

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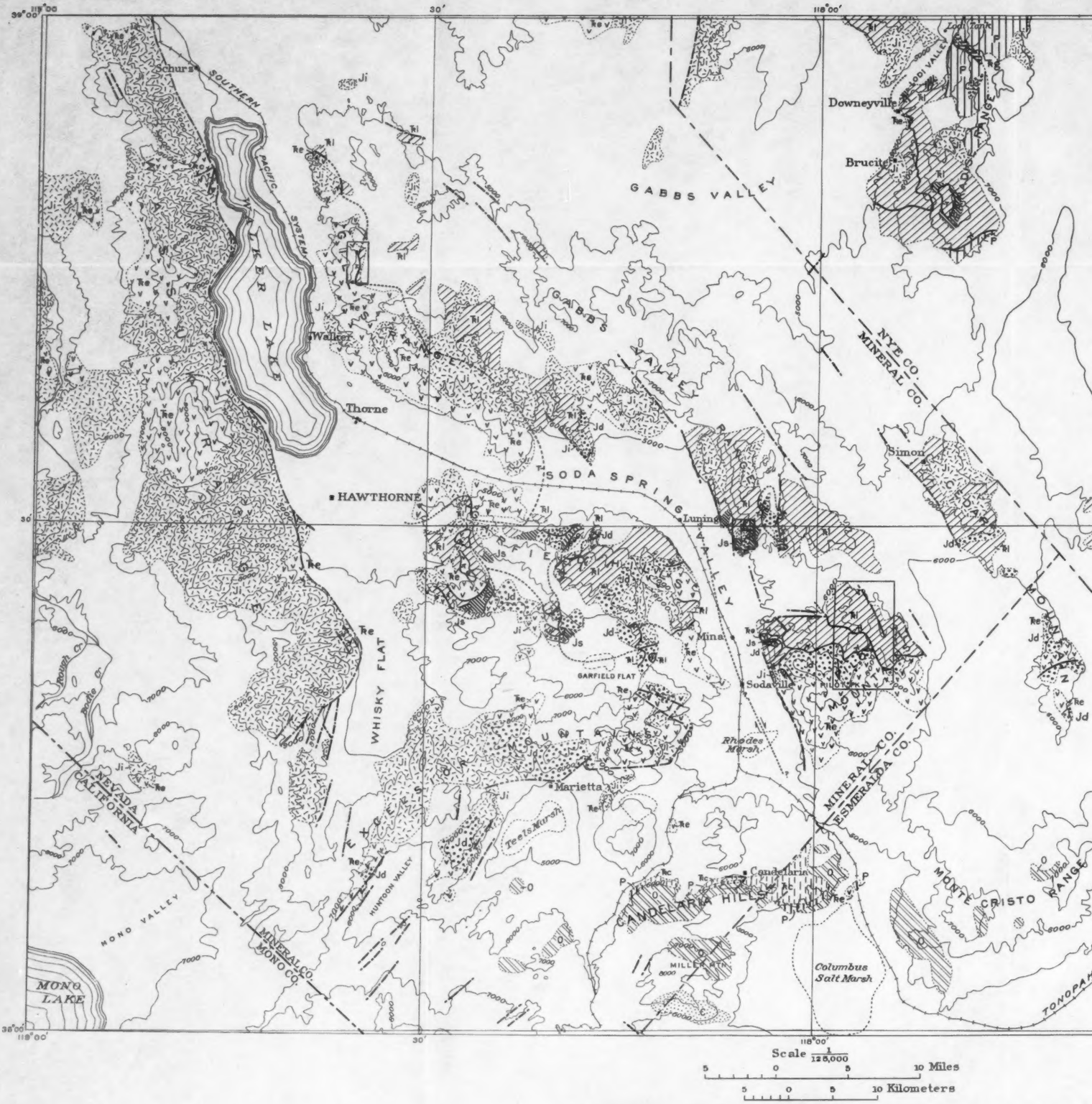
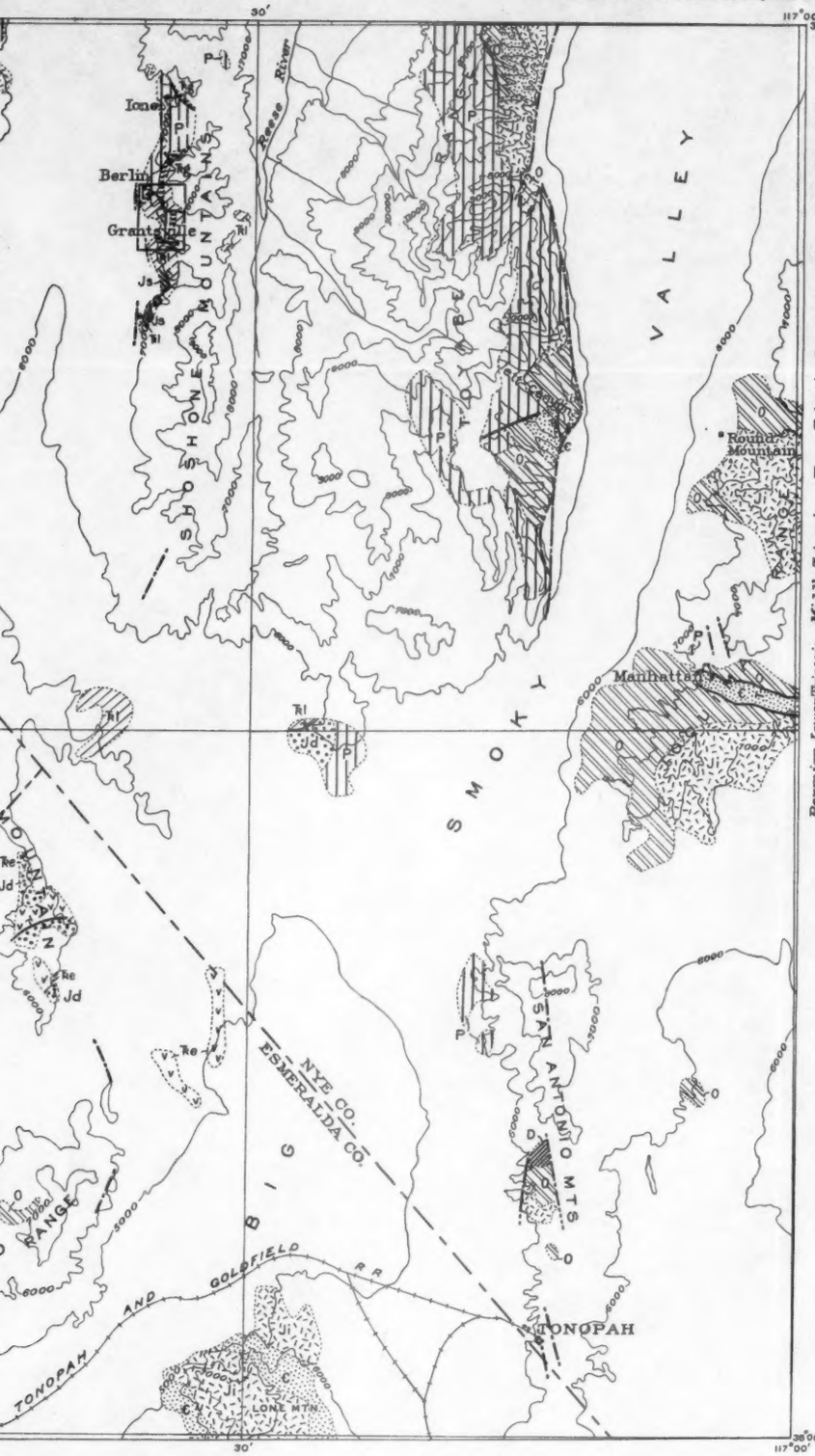


MAP SHOWING DISTRIBUTION OF PRE-TERTIARY FORMATIONS OF THE HAWTHORNE AND TONOPAH QUADRANGLES.

Boundaries generalized and minor areas omitted or exaggerated. Areas in small rectangles are shown in larger scale in Figure 2 and Plates 2, 3 and 4. Geology by H. G. Ferguson, S. W. Muller, S. H. Cathcart, and W. F. Foshag.



EXPLANATION

- Intrusives (Granodiorite, quartz monzonite, diorite, aplite etc.) — Upper Jurassic(?) — JURASSIC(?)
- Dunlap formation (Lava, tuff, conglomerate, conglomerate, and sandstone) — Upper Jurassic — JURASSIC
- Sunrise formation, including Gabbs formation (Upper Triassic) (Limestone and shale) — Lower Jurassic — JURASSIC
- Luning formation (Dolomite, limestone, slate, and conglomerate) — Upper Triassic — TRIASSIC
- Grantsville formation (Limestone, slate, and conglomerate) — Middle Triassic — TRIASSIC
- Excelsior formation (Lava, tuff, and chert) — TRIASSIC
- Candelaria formation (Slate, sandstone, limestone, and conglomerate) — TRIASSIC
- (Lava, tuff, chert, dolomite, sandstone, and conglomerate) — Permian Lower Triassic — PERMIAN
- (Limestone and slate) — DEVONIAN
- (Schist, slate, limestone, and chert) — ORDOVICIAN
- Quartzite, schist, dolomite, and limestone — CAMBRIAN

Pre-Tertiary thrust faults
T, overthrust side

Pre-Tertiary flaws and normal faults
Arrows indicate relative horizontal movement. Faults dotted where concealed by later deposits or obliterated by later intrusion

Tertiary and Quaternary normal faults
(In large part concealed by later deposits)

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MESOZOIC STRATIGRAPHY OF THE HAWTHORNE AND TONOPAH QUADRANGLES, NEVADA

BY SIEMON WM. MULLER AND HENRY G. FERGUSON

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1. Map showing distribution of pre-Tertiary formations of the Hawthorne and Tonopah quadrangles 1573
2. Geologic map of part of Shoshone Mountains between Union Canyon and Grantsville Canyon 1592
3. Geologic map of north-central part of Pilot Mountains 1596
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ABSTRACT

Mesozoic rocks of the Hawthorne and Tonopah quadrangles include Triassic and Lower Jurassic formations that occur in isolated areas in the different mountain ranges of the area and have a maximum thickness of 30,000 feet. The mapped formations are (from top to bottom):

	Feet
Dunlap formation—Lower Jurassic—Coarse clastic sedimentary rocks and volcanics	5000
Unconformity (Local conformity)	
Sunrise formation—Lower Jurassic—Marine shales and limestones	1240
Gradational contact	
Gabbs formation—Upper Triassic—Marine shales and limestones	420
Conformity	
Luning formation—Upper Triassic—Limestones, dolomites, and shales, some conglomerate	8000±
Unconformity	
Grantsville formation—Middle Triassic—Limestone, argillite, and conglomerate	800
Excelsior formation—Middle Triassic—Cherts, altered volcanic rocks with lenses of marine shales and limestone. May be in part equivalent to the Grantsville formation	12,000
Contact not exposed	
Candelaria formation—Lower Triassic—marine shales and limestones with tuffaceous sandstones and shales	3000
Unconformity	
Permian and older rocks	

The lithology, areal extent, fauna, and correlation of each formation is described.

Rich fossil collections from the area furnished much material new to this continent. The *Claraia* and *Proptychites* assemblages from the Lower Triassic Candelaria formation are earlier than the *Meekoceras* fauna, which has been heretofore known as the oldest Triassic ammonite fauna of this continent. The fossils from the Grantsville and the Excelsior formations, although differing from one another, both show some affinity to the European Muschelkalk fauna. Reef-building corals

from the Luning formation are correlated with the Karnic of the European section and not the Noric as was previously held by other writers. A new Gümbelites fauna above the *Tropites* assemblage in the Luning probably represents a transitional Karnic-Noric fauna. Three zonal assemblages were recognized in the Gabbs formation: Zone of *Sagenites giebeli*, zone of *Pinacoceras metternichi*, and the zone of *Pteria contorta* (or *Choristoceras marshi*). All three are new to North America. The *Choristoceras* fauna is of particular interest, as marine rocks equivalent to the Rhaetic of the European section heretofore have not been recognized in North America. These Rhaetic beds are overlain with a gradational contact by the lowermost Jurassic beds containing the ammonites *Psiloceras, Waehneroceras,* and *Euphyllites.* These ammonites likewise have not been hitherto recognized on this continent. The gradational contact between the Triassic and Jurassic systems in western North America permits a revision of the previously held views that "nowhere on the continent is there any latest Triassic or earliest Jurassic sediment."

The sudden change in most of the area from marine deposition of the Sunrise and older formations to subaerial and volcanic debris of the Dunlap is ascribed to acute diastrophism during the latter part of the early Jurassic.

INTRODUCTION

LOCATION AND CHARACTER OF AREA

The Hawthorne and Tonopah quadrangles (Fig. 1) are situated in west-central Nevada, between north latitudes 38° and 39° and between west longitudes 117° and 119°; the area covers over 7700 square miles—slightly smaller than the State of Massachusetts.

The region is readily accessible. The principal towns—Tonopah, Mina, and Hawthorne—have rail and bus connection with Reno, and a number of improved highways, together with numerous desert roads, make it possible to reach most of the area by automobile.

The Mesozoic rocks of the two quadrangles, exclusive of the granitic intrusions, are confined to mountain ranges in the northern and eastern parts of the Hawthorne quadrangle east of Walker Lake and in the northwestern part of the Tonopah quadrangle, as shown in Plate 1. Even in these mountain ranges the unmetamorphosed Mesozoic rocks are confined to rather small areas, much of the surface being underlain by Tertiary rocks, valley fill, and granitic intrusions.

FIELD WORK AND ACKNOWLEDGMENTS

The field work in the area was begun by H. G. Ferguson and S. H. Cathcart in 1922 and continued during the field seasons of 1923 and 1924; in the course of the preliminary areal mapping much help was received from the following assistants: W. F. Foshag, Q. D. Singewald, and H. H. Chen.

During the early reconnaissance, T. W. Stanton twice visited the region and was of invaluable aid to the writers through his advice and interpretation in the laboratory as well as in the field.

Work on the Mesozoic stratigraphy and paleontology was started by S. W. Muller in 1927, and since 1928 both Ferguson and Muller, sepa-

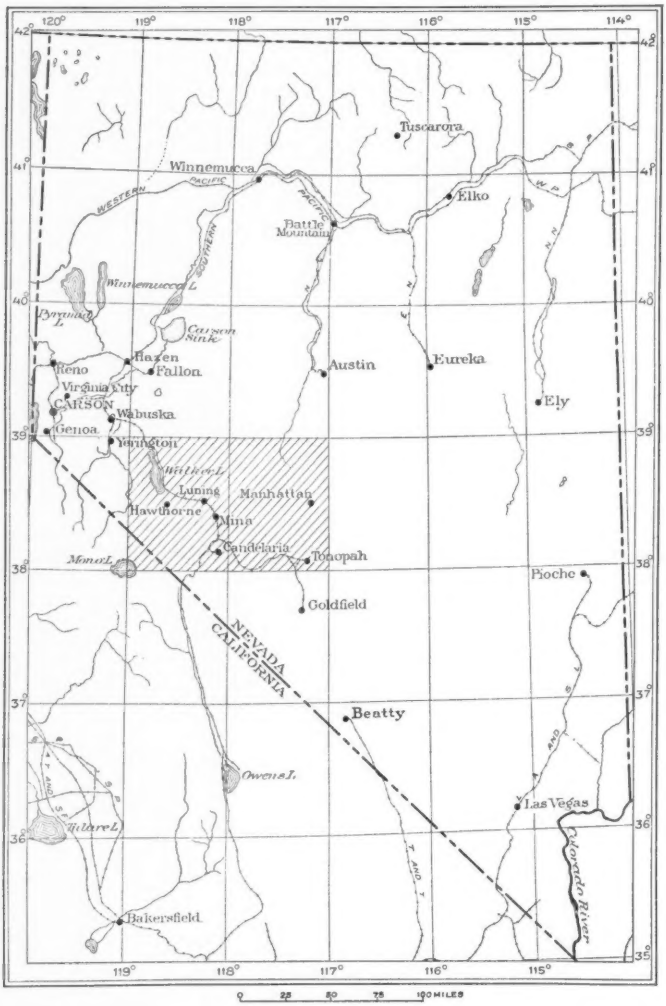


FIGURE 1.—*Map showing location of Hawthorne and Tonopah quadrangles, Nevada*

rately and jointly, have been engaged from time to time in detailed geologic mapping of the more complex areas.

Thanks are also due M. E. Souza, H. E. Wheeler, J. F. McAllister,

and J. D. Barksdale, who helped the writers in geologic mapping of various areas. The writers wish to express, in particular, their indebtedness to the late Dr. James Perrin Smith, whose personal interest in the problem, helpful suggestions, and constructive criticism freely offered at all times in his usual informal manner were of immeasurable value.

The work was carried out under the auspices of the United States Geological Survey and, in part, with the assistance of the Stanford University Research Council, whose financial aid is gratefully acknowledged. Thanks are also due to The Geological Society of America, whose generous grant from the Penrose Bequest made it possible for both authors to do field work during the season of 1934.

In the course of this work a large amount of fossil material was collected. This will be described in a paper now in preparation.

FIELD METHODS

When the project was started by Ferguson in 1922 it was planned to prepare a reconnaissance geologic map using the available one-degree topographic sheets as the base. A map of this scale, however, was inadequate to show geologic detail, and it later became necessary to remap a number of important and critical areas on a much larger scale. The detailed geologic mapping was begun by Muller in 1928 and carried out with differing degrees of speed and accuracy. For the New York Canyon area, for example, a geologic map (Pl. 4) was made by Muller on the scale of 1:6000 from an accurately chained base and with a carefully prepared triangulation network. For other areas mapped in detail by both writers, more rapid and less accurate methods were used. For some a base was measured with a telescopic alidade and a stadia rod, and a certain amount of control was established through graphic triangulation. For several areas a paced base sufficed. Topography was sketched by compass intersections of prominent points with 100-foot contours controlled by aneroid.

PREVIOUS WORK

Previous knowledge of the Triassic and Jurassic paleontology and stratigraphy of the Hawthorne and Tonopah quadrangles is here summarized and arranged in chronologic order. The notes pertain only to those parts of the article which deal with the problem of the present paper. The results of the present investigation, wherever they differ from conclusions of previous authors, are also alluded to briefly.

The earliest account of the Mesozoic rocks in the Hawthorne and Tonopah quadrangles was given by Whitney (1866, p. 268). He states that Triassic fossils have been found in the Volcano district and that

> "there are indications of the existence of fossiliferous rocks occupying a higher position than the Trias, and perhaps of Liassic Age."

The fossils collected in the Volcano district by J. S. Clayton and Dr. A. Blatchley were described and figured by Gabb (1870, p. 5-18). Gabb did not visit the locality himself and in determining the ages of some of the less diagnostic fossils he had to rely solely on the lithologic character of the matrix. It is therefore not surprising that some of the Jurassic forms were dated by him as Triassic, and vice versa. The fauna described by Gabb will be discussed at greater length in the forthcoming paleontological report; it will suffice at this time only to point out that Gabb definitely established the presence of both Triassic and Jurassic rocks in the Volcano district (Gabbs Valley Range).

To the list of Jurassic fossils from this area Hyatt (1892, p. 411) added the new species, *Vermiceras crossmani*, which he regarded as a lower Liassic form "of undeniable European facies".

Hyatt (1894, p. 418) later described another new species, *Coroniceras claytoni*, and, in giving a general summary of the Triassic and Jurassic succession in the Western States, pointed out that the upper Lias as well as the lower Lias is present in the Gabbs Valley Range. Among the Triassic fossils in the Whitney collection from the Volcano district Hyatt reported a specimen of *Arcestes* and a slab of *Daonella* "having the same catalogue number and apparently from the same locality". During the past 10 years of extensive collecting by the present writers not a single specimen of *Daonella*, which is a typical Middle Triassic genus, had been found in the Gabbs Valley Range. There is thus every reason to believe that the *Daonella* slab in the Whitney collection probably came from some other area and was wrongly labelled. This unfortunate circumstance explains why the common Upper Triassic *Arcestes "nevadanus" ("Joannites nevadanus")* was described by Hyatt and Smith (1904, p. 76) as new and was incorrectly assigned by them to the Middle Triassic.

Turner's (1902, p. 267) brief account of the Pilot Mountains gives the erroneous idea that the entire range is made up of Jurassic rocks. The meager collection of fossils of no diagnostic value reported by Turner as Jurassic more likely belongs to a late Triassic fauna widely distributed in this range. It is true, however, that the sandstones and conglomerates (Dunlap formation) that constitute a part of the Pilot Mountains section are of early, although probably not the earliest, Jurassic age. The geologic cross section of the Pilot Mountains shown by Turner is highly diagrammatic and does not represent the actual relations of the rocks of that range.

Spurr (1905, p. 99-113) gives a brief description of Triassic sedimentary and volcanic rocks in the Paradise Range ("Ellsworth Range"), Pilot Mountains, and Excelsior Range. He correlates these rocks with the Star Peak and the Koipato formations of the West Humboldt Range

in northwestern Nevada. The fossiliferous limestones at the north end of the Pilot Mountains reported by Spurr as Jurassic are now known to be of late Triassic age. The present writers' views on the correlation of the Star Peak and the Koipato formations with the rocks in the Hawthorne and Tonopah quadrangles is shown in Table 3.

Hyatt and Smith (1905, p. 76) described the ammonite *Joannites nevadanus* Hyatt and Smith from the Volcano district and referred it to the upper Muschelkalk (upper Middle Triassic) of the European section. During the present investigation no Middle Triassic faunas have been found in the Volcano Peak region, whereas *"Joannites nevadanus"* was found to be one of the most common forms in the *Pinacoceras metternichi* zone of the Upper Triassic Gabbs formation. The species *J. nevadanus,* later transferred by Smith (1914, p. 44) to the genus *Arcestes,* is now believed to be conspecific with *Arcestes intuslabiatus* Mojsisovics from the upper Noric of the Alpine region.

The reef-building corals from the north end of Pilot Mountains were first dated by J. P. Smith as Jurassic (Spurr, 1905, p. 104). In 1912 he assigned these corals to the interregional coral zone of early Noric age. It now appears that these corals are of Karnic age (Muller, 1936, p. 202-208).

Hill (1915) recorded the occurrence of Triassic rocks in the Gabbs Valley Range, the Garfield Hills, and the Excelsior Mountains. He correlated the sedimentary rocks with the Star Peak formation of the West Humboldt Range and the volcanic rocks with the Koipato formation; he also recognized the tuffaceous nature of the cherts in the Excelsior formation.

Knopf (1922) briefly described the geology of the northern part of the Cedar Mountain range. The age of the limestone found there was considered to be Middle Triassic, but the present studies show it to be Upper Triassic. No surficial volcanic rocks of Triassic age mentioned by Knopf were found in this area by the present writers.

In describing the geology near Volcano Peak, Clark (1922, p. 8, 9) states that the Mesozoic rocks of the region may be of two ages but that the

> "faunas are too meager and poorly preserved to prove conclusively that distinct Triassic and Jurassic formations are represented in the district."

Ferguson and Cathcart (1924) in a brief note dealt chiefly with the structures of the area and gave only a brief and generalized statement pertaining to the stratigraphic sequence of the Mesozoic rocks.

Stanton (1926) recognized in the Triassic of this area two new assemblages of marine invertebrates which he dated as Noric. The younger of these two assemblages is characterized by the presence of *Cladiscites*

aff. *C. tornatus, Rhabdoceras,* and *Rhacophyllites,* whereas the other contains such forms as *Myophoria, Ostrea (Alectryonia),* and *"Pinnigena".* Present studies show that the older assemblage corresponds in age to the Karnic stage of the Upper Triassic of Europe instead of to the Noric.

Smith (1927) described several genera and species of reef-building corals from Dunlap Canyon in the Pilot Mountains, assigning them to the interregional coral zone of early Noric age. These corals are now correlated with those of the Karnic stage of the European section.

Foshag (1927, p. 115) published a brief account of the stratigraphy of the Pilot Mountains; he presented a geologic map of the area. This map and the accompanying description is based on the reconnaissance work of Ferguson and Cathcart in their areal mapping of the Hawthorne and Tonopah quadrangles. At that time it was supposed that the conglomerates and sandstones (Dunlap formation) were intermediate in stratigraphic position between the cherts and greenstones (Excelsior formation) and the argillites and limestones (Luning formation); the thrust fault—the key to the Pilot Mountains structure—had not been recognized in the reconnaissance mapping.

Crickmay (1931) published an admirable compilation of data on Jurassic stratigraphy and paleontology of North America. The parts (p. 20) which deal with the area of the present report are, however, based on publications which are decidedly out of date and are not free from errors. Referring to the present writers' collections of fossils from the Lower Jurassic Sunrise formation, Crickmay states that the Volcano district faunas range "from Schlotheimian to Hildoceratan ages"—*i. e.,* from the Upper Hettangian to the middle Toarcian of the European section. In the present report the Sunrise formation fauna is regarded as ranging from the base of the Hettangian ("Psiloceratan") to the lower part of Pliensbachian ("Deroceratan") of the European standard column.

Muller (1936) presented field evidence which shows that the Dunlap Canyon corals are not a part of the lower Noric "Interregional Coral Zone," as was thought by Smith, but are correlated with the Karnic stage of the European section.

Muller and Ferguson (1936), in a preliminary report on the Mesozoic stratigraphy of the Hawthorne and Tonopah quadrangles, introduced the formational names used in the present report and briefly noted the lithology and characteristic fossils of each formation. A new formational unit of Middle Triassic age (Grantsville formation) has since been added to the Mesozoic sequence of the area and is described for the first time in the present article.

Ferguson and Muller (1936) discussed briefly the structural relations of the Mesozoic formations of the region with no additional contribution to the stratigraphic succession.

A more complete list of publications consulted in the preparation of this article will be found at the end of this paper.

GENERAL STRATIGRAPHY

The Mesozoic rocks of the Hawthorne and Tonopah quadrangles, with a maximum thickness of about 30,000 feet, are predominantly marine clastic sediments, cherts, and limestones, with considerable altered pyroclastic rocks and lavas. They range from Triassic to Jurassic (Pl. 1).

The profound crustal deformation and igneous activity which this region has undergone since the deposition of the Mesozoic sediments has in many places obscured the true relationships of the formations. The folded and faulted Mesozoic rocks occur in isolated patches in an extremely irregular mosaic interspersed with areas of granites, Tertiary volcanic rocks, and alluvium. At no single locality within the two quadrangles is there a complete and undisturbed succession of all the Mesozoic formations, although a number of sections, representing substantial parts of the sequence, were found.

The formations which constitute the Mesozoic section in this area are shown in Table 1.

TRIASSIC SYSTEM

CANDELARIA FORMATION (LOWER TRIASSIC)

Definition and lithologic character.—The Candelaria formation is the lowest Mesozoic unit in the area. The type locality of this formation is about 2 miles southeast of the Candelaria mining camp, in the low hills which are indicated on Plate 1 as the Candelaria Hills.

The Candelaria formation rests with slightly angular unconformity upon the Permian and with markedly angular unconformity upon the Ordovician. The intermittent absence of Permian rocks along the Candelaria contact indicates some deformation and erosion preceding Candelaria deposition.

The formation is more than 3000 feet thick and consists essentially of greenish-brown shales, sandy shales and sandstones, some of which have a tuffaceous aspect, and in places thin layers and lenses of limestone. The basal conglomerate is not everywhere present but wherever observed is only about 1½ feet thick.

A traverse across the exposed portion of the formation at the type

TABLE 1.—*Mesozoic section in Hawthorne and Tonopah quadrangles, Nevada*

Series	Standard European Stages	Formations	Est. Max. Thickness (Feet)
		Erosion surface capped by Tertiary volcanic rocks and alluvium Intrusions of granitic and subordinate basic rocks allied to the Sierra Nevada batholith.	
.....	.....	.....	
Lower Jurassic..	Lias.....	DUNLAP FORMATION.....	5000±
		Unconformity (Local conformity).....	
		SUNRISE FORMATION.....	1240
.....	.....	Gradational contact.....	
	Rhaetic..... Noric.....	GABBS FORMATION.....	420
Upper Triassic..		Conformity.....	
	Karnic.....	LUNING FORMATION.....	8000±
.....	.....	Unconformity.....	
	Ladinic.....?	— GRANTSVILLE FORMATION..	800
Middle Triassic.		..... ? ..... ?	
	Anisic.....	EXCELSIOR FORMATION..... (In part probably correlative of Grantsville)	12,000±
.....	.....	Contact not exposed.....	
Lower Triassic..	Scythic.....	CANDELARIA FORMATION.....	3225±
.....	.....	Unconformity.....	
Permian and older		(Permian rests unconformably on Ordovician, but there is a small isolated area of Devonian limestone and slate in the San Antonio Range, north of Tonopah.)	

locality southeast of Candelaria showed the following sequence from top to bottom:

	Feet
(5) Fine flaky shales with a few beds of impure iron-stained limestone, grading downward into fine-grained siliceous sandstone	1000
(4) Predominantly massive sandstone beds of various shades of brown. Some beds probably contain tuffaceous material	1000
(3) Olive-green to greenish-brown slaty shales with platy or splintery fracture, locally siliceous, and generally closely jointed. Small lenticular masses of limestone, locally veined and in part replaced by barite	1000
(2) Fossiliferous, dark bituminous shales, some layers containing flat ellipsoidal concretions and a few thin beds of dark impure limestone, purplish on weathered surface	75

	Feet
(1) Purplish-gray to black micaceous shales, thinly laminated, with occasional layers of sandstone. Several beds of gray massive limestone 1 or 2 feet thick occur at intervals of 10 to 20 feet. Basal conglomerate up to 1½ feet thick, not everywhere present	150
Total thickness exposed	3225+

The total thickness of the formation is not known, as near the top it is covered by desert wash, and the contact with the Excelsior formation is also everywhere concealed. The nearest outcrop of Excelsior is about 4 miles north of the northernmost exposure of the Candelaria formation.

Eastward from the type locality the shales pass into dark, seemingly tuffaceous sandstone interbedded with conglomerate. Individual conglomerate beds average from 1 to 2 feet in thickness and appear to be lenticular, although several beds can be traced for over 100 feet. The pebbles in the conglomerate are almost entirely of chert, ranging through various shades of dark gray to black, with occasionally one of jasper red. The pebbles are rounded to angular and range in size from a fraction of an inch to over a foot in diameter. The matrix is commonly dark greenish-gray fine to coarse sandstone and occasionally even grit of comminuted angular fragments of chert.

In the upper part of the Candelaria formation several more or less lenticular masses of partly silicified dolomite form prominent outcrops. These rugged masses are roughly aligned with the strike of the beds but here and there appear to cross the bedding. Field evidence indicates that at least some of these dolomites are the product of alteration surrounding small basic intrusives, for cores of serpentine are completely enveloped by dolomite.

Fossils and correlation.—The only fossils found in the formation came from the 75-foot thickness of bituminous shales and limestones about 150 feet above the base of the formation. The collection from the lower 25 feet of this fossiliferous zone consists of several fragments of indeterminate ammonites and many specimens of the pelecypod genus *Claraia*. The three species in this collection, in the order of their abundance, are: *Claraia stachei* (Bittner), *C. aurita* (Hauer), and *C. clarai* (Emmrich).

In southern Tyrol the species *C. clarai* is restricted to the lower part of the Seis member of the Werfen formation. The species *C. aurita*, although occurring in both the lower and the upper division of the Seis, does not extend into the overlying Campil member of the Werfen formation. The Indian species *C. griesbachi* Bittner, a form somewhat similar to *C. aurita*, according to Diener (1909, p. 165)

"is the leading species of the Otoceras beds, both in Painkhanda and Spiti. In the Shalshal cliff section Noetling found its main layer on the top of the limestone bed

containing *Otoceras Woodwardi*. In Spiti it is most common in the beds containing *Ophiceras* and immediately overlying the ferruginous limestone with *Otoceras*."

Spath (1930, p. 46-47) described *Claraia stachei* (Bittner) from Greenland, where he found it in the *Ophiceras* and the *Proptychites* beds. In a more recent report on the Lower Triassic of Greenland Spath (1935) shows the species *C. stachei* ranging from the upper *Glyptophiceras* beds to and including the *Proptychites* beds or through the upper "Otoceratan" and the lower "Gyronitan".

Claraia clarai also has been reported recently from Eastern Transbaikalia (Kiparisova, 1932, p. 22, 23, 29).

The occurrence of these three species of *Claraia* in the Lower Triassic of North America makes them world wide in distribution, and with their relatively restricted stratigraphic range they become valuable for correlation. In North America this fauna represents a much lower part of the lower Triassic than the *Meekoceras* zone of Idaho and California (Smith, 1932, p. 15), which has been heretofore the earliest Triassic ammonite fauna known from this continent.

As to the interregional correlation, this lower fauna from the Candelaria formation may be regarded as the time equivalent of the assemblage from the lower part of the Seis member of the Werfen formation in Tyrol and possibly also, although only in part, of faunas from the upper "Otoceratan" or lower "Gyronitan" beds of the Scythic stage in Greenland and India (Spath, 1935, p. 102). In Smith's classification of the Triassic this fauna probably corresponds to the upper part of the *Otoceras* zone.

It is, of course, realized that this correlation with India and Greenland is based on a rather meager fauna and perhaps should be regarded as tentative; nevertheless, the general arrangement appears to be correct. It should be noted that Diener (1912, p. 65) also regarded the *Claraia clarai*-bearing beds of the Tyrol as equivalent, at least in part, to the *Otoceras* beds of India. Further evidence in support of this correlation is furnished by the next higher fauna from the Candelaria formation, obtained from the beds approximately 40 to 50 feet above the *Claraia*-bearing beds and containing the following species: *Hedenstroemia (Clypites)* cf. *H. (C.) evolvens* Waagen, *Meekoceras* cf. *M. lilangense* Krafft, *M.* cf. *M. tenuistriatum* Krafft, *Proptychites* cf. *P. ammonoides* Waagen, *P.* cf. *P. trilobatus* Waagen, *Grypoceras* cf. *G. brahmanicum* (Griesbach), *Gyrolepis?* sp.

This entire assemblage also appears to be new to the North American continent. It is most nearly related to the early Triassic fauna from India.

Meekoceras lilangense Krafft, *Meekoceras tenuistriatum* Krafft, and *Grypoceras brahmanicum* (Griesbach) have been previously known only from the so-called "*Meekoceras* beds" of the Himalaya Mountains,[1] whereas *Proptychites ammonoides* Waagen, *P. trilobatus* Waagen, and *Hedenstroemia (Clypites) evolvens* Waagen were recorded only from the Ceratite marls of Salt Range. There has been as yet no field evidence found to suggest that this mixed association of species in Nevada might be ascribed to a condensation of heterochronous faunal elements due to nondeposition or a nonsequence. The Nevada assemblage therefore, at least for the present, is regarded as equivalent to a part of the Ceratite marls of Salt Range and a part of the "*Meekoceras* beds" in the Himalaya Mountains.

The correlation of the Lower Triassic rocks of Salt Range with those of the Himalaya Mountains has been debated for some time. The difficulty in the correlation of the rocks of these two areas was ascribed by Diener (1912, p. 35) to:

"the special development, which the Lower Triassic faunae have taken in those two regions" . . . ". . . a correlation of the minor subdivisions cannot be made except in a most general way, and always with the mental reservation that their lower and upper boundaries do not strictly coincide."

Spath (1934, p. 27) in his recent subdivision of the Lower Triassic places the "*Meekoceras* beds" of the Himalaya Mountains at the base of the Owenitan division. The Ceratite marls of the Salt Range is, however, shown by him as ranging from Gyronitan to Flemingitan, which is at variance with the correlation established by Diener (1912, p. 55).

The *Proptychites* fauna of Nevada, as already mentioned, contains a mixture of faunal elements some of which heretofore have been regarded as distinctive of the Salt Range province, whereas others have been known only in the Himalayan Triassic. The *Proptychites* horizon of Nevada, therefore, may be looked upon as a common stratigraphic level for the correlation of strata in these two provinces in India, and the correlation thus established places the Salt Range Ceratite marls, at least in part, as equivalent to the "*Meekoceras* beds" of the Himalaya Mountains—a relationship which is in a closer agreement with the scheme of Diener than with that of Spath.

In the table of the *Subdivisions of the Lower Trias* by Spath (1934, p. 27) this new assemblage from Nevada can be tentatively assigned to the "Gyronitan", while in Smith's (1932, p. 15) terminology it would fall in the *Genodiscus* zone.

[1] The "*Meekoceras* beds" of India (and Madagascar), as has been shown by Smith (1932, p. 15), are not equivalent of the *Meekoceras* zone of California and Idaho. Smith correlated the "*Meekoceras* beds" of the Himalayan province with the *Genodiscus* zone which is the next zone below the *Meekoceras* zone.

As the *Claraia* and *Proptychites* faunas have not been recognized elsewhere in North America, the Candelaria formation cannot be directly correlated with any of the known Lower Triassic units in this continent. The Woodside shale which underlies the *Meekoceras* zone in northern

TABLE 2.—*Stratigraphic distribution of fossils of the Candelaria formation*

	Lower Triassic						
Smith 1932	*Otoceras* zone	*Genodiscus* zone	*Meekoceras* zone		*Tirolites* zone	*Columbites* zone	
Spath 1935	Otoceratan	Gyronitan	Flemingitan	Owenitan	Columbitan		Prohungaritan
(Faunas previously reported from North America)			x	x	x	x	?
Meekoceras lilangense Krafft		x					
M. tenuistriatum Krafft		x					
Proptychites ammonoides Waagen		x					
P. trilobatus Waagen		x					
Grypoceras cf. *G. brahmanicum* (Griesbach)		x					
Claraia clarai (Emmrich)	x						
C. stachei (Bittner)	x						
C. aurita (Hauer)	x						

Utah and Idaho (Table 3) may well be the equivalent of the Candelaria, but definite proof for this is still lacking. The Moenkopi formation may also be, although perhaps only in part, equivalent to the Candelaria formation. Here too, no direct correlation is possible, as the lenticular marine members in the Moenkopi formation show a different faunal and lithologic facies. It is even less feasible, at least for the present, to correlate the Candelaria formation with either the Dinwoody or the Chugwater formations of Wyoming.

EXCELSIOR FORMATION (MIDDLE TRIASSIC)

Definition and lithologic character.—The Excelsior formation, named from the Excelsior Mountains, about 6 miles southwest of Mina, consists dominantly of volcanic rocks and sedimentary rocks composed for the

most part of water-laid tuffs and cherts. Although no sequence has been worked out for this very heterogeneous formation as a whole, or for any individual mountain range, the thickest sections of cherts and other sediments were seen in the southwestern part, in the Pilot Mountains, southeastern Garfield Hills, and Excelsior Mountains. The total thickness of the formation is estimated at 12,000 feet.

As no detailed petrographic study of the volcanic rocks has yet been made, it seems advisable to use the noncommittal terms greenstone and felsite respectively for the more basic types of lavas (largely andesitic) and for those which show free quartz (including rhyolite and quartz latite). Most of the thin sections of the volcanic rocks studied are from mining districts. These show the effects of widespread albitization, presumably a hydrothermal process. Chloritized rock is widespread, and there is also much epidote, particularly along joint planes. This widespread occurrence of epidote is a useful criterion for distinguishing isolated outcrops of the Excelsior lavas from the more highly altered Tertiary lavas.

Basic intrusives, including masses and dikes of diabase, gabbro, and diorite, are much more abundant in the Excelsior than in the succeeding formations, and probably the larger proportion of these intrusives are of Excelsior age; as diorite and gabbro also cut the younger formations the proportion of basic intrusive rocks of pre-Luning age is indeterminable. Serpentine dikes, which cut the Lower Triassic Candelaria formation, have not been found in the Excelsior or higher formations and may belong to this earlier period of volcanism. Siliceous intrusives other than small dikes of the same general type as the felsites appear to be lacking or, if present, cannot be definitely segregated from the later granitic intrusives.

In the southern part of the Pilot Mountains and neighboring parts of the eastern Garfield Hills and Excelsior Mountains there is massive dark chert estimated to exceed 10,000 feet in thickness, with individual layers varying in thickness from several inches to about 2 feet. Under the microscope the presence of minute strands of devitrified glass suggests that the chert is of tuffaceous origin (Hill, 1915, p. 20; Foshag, 1927, p. 115). The chert grades into and is interbedded with recognizable andesite tuffs, tuffaceous slates, and dark sandstones. To date no fossils have been found in sedimentary rocks of this type. In the Excelsior Mountains the sedimentary member is overlain by thick flows of felsite.

The contact of the Excelsior formation with the Candelaria formation is not exposed but, from the areal distribution of the two formations and also on the basis of the fossils found in the Gillis Range in the

northwestern part of the area, it is concluded that the Excelsior formation overlies the Candelaria. As the Gillis Range (Fig. 2) is the only locality which yielded identifiable Middle Triassic fossils, the age

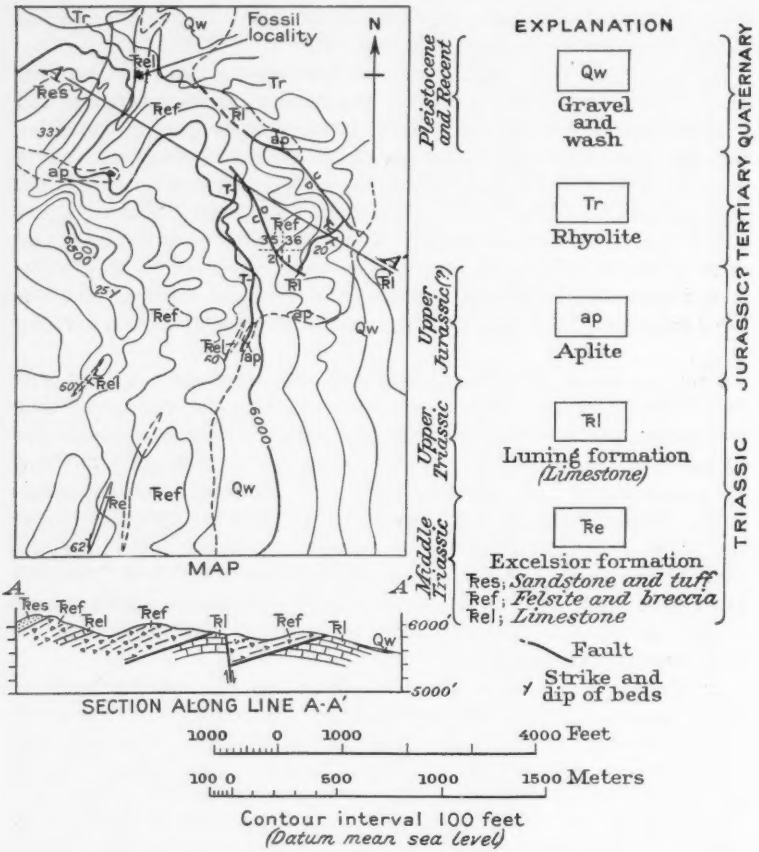


FIGURE 2.—*Geologic map of area near Gillis Canyon, Gillis Range*

assignment of the formation in other isolated areas is doubtful, particularly as similar lavas and tuffs are also known in the Permian section of the Toyabe Range in the eastern part of the area.

The upper stratigraphic limit of the Excelsior formation likewise cannot be accurately determined, for the formation is overlain with

marked angular unconformity by the Luning formation of late Triassic age.

Fossils and age.—The only evidence for dating the volcanic rocks of the Excelsior formation was obtained in the Gillis Range in the vicinity of Gillis (Wildhorse) Canyon in the lenticular limestones and shales interbedded with lavas and volcanic breccias. Similar lenses of limestone and shale interbedded with volcanic rocks were observed in the eastern part of the Garfield Hills and in the Wassuk Range, but at these localities the sedimentary rocks are too highly metamorphosed to yield determinable fossils.

The fossils collected from the Excelsior formation in the Gillis Range, although poorly preserved, were tentatively identified as:

Spiriferina cf. *S. fragilis* (Schlotheim)
S. sp.
Lima cf. *L. lineata* (Schlotheim)
Hörnesia aff. *H. socialis* (Schlotheim)
Pleuromya mactroides (Schlotheim) of Goldfuss
Gervilleia (three species)
Pecten aff. *P. tirolicus* (Wittenburg)
Myophoria sp.
Pleurophorus sp.

Of these forms *Spiriferina fragilis* is known to occur in the Muschelkalk (Middle Triassic) beds of Germany, the Alps, Asia Minor, and New Caledonia. *Lima lineata* (Schlotheim) which resembles very closely the form from the Excelsior formation is also a Muschelkalk (Middle Triassic) species. The strongly twisted *Hörnesia socialis* is known only from the lower part of the German Muschelkalk. *Pleuromya mactroides* (Schlotheim) in Europe ranges from Röth (uppermost Buntersandstein) to Keuper, although the particular form (Goldfuss, 1862, pl. 154, fig. 1) which compares most closely with the Nevada specimen comes from the Muschelkalk beds of Germany. Associated with these forms is a *Pecten*, which appears to be related to the Campiler (Lower Triassic) species *P. tirolicus* from the Tyrol. *Gervilleia, Pleurophorus*, and *Myophoria* have not yet been specifically determined.

On the whole the resemblance between the Gillis Range fauna and the Muschelkalk faunas suggests early Middle Triassic as the age of the Excelsior formation, although the possibility is not excluded that it may be of latest Lower Triassic age.

Correlation.—The faunal assemblage from the Excelsior formation has some affinity with the faunas which are intermediate between the typical Muschelkalk of Germany and the Middle Trias of the Alpine province. A somewhat similar fauna is known from the Silesian region, where, as is generally believed, a passage existed between the Tethyan

Sea to the south and the Muschelkalk "Binnen" Sea northwest of the Bohemian massif.

Middle Triassic faunas of the Excelsior facies have not been heretofore recognized on this continent, and for this reason the Excelsior cannot be directly correlated on the basis of fauna with other Middle Triassic formations in North America.

Volcanic rocks of supposedly Triassic age and in a general way resembling those of the Excelsior formation have been recognized at a number of localities outside the area of the Hawthorne and Tonopah quadrangles. It is believed, however, that the correlation of these rocks on the basis of lithology alone cannot be carried out with any degree of certainty, although in the past there has been a strong tendency on the part of various writers to place an undue reliance on lithologic similarity in correlating these volcanic sections.

It is proposed to review briefly some of these occurrences of volcanic rocks and to comment on the validity of the previously established correlations.

The nearest area of Triassic volcanic rocks outside the Hawthorne and Tonopah quadrangles is at Yerington, about 12 miles west of the northwestern corner of the Hawthorne sheet. These rocks were described by Knopf (1918) who assigned them to the Middle Triassic; they may tentatively be regarded as the equivalent of the Excelsior formation.

The volcanic rocks of the Koipato formation in West Humboldt Range were originally placed by King (1878, p. 268-273) in the Lower Triassic, but Knopf (1924, p. 9, 13) in a detailed account of the stratigraphy of the Koipato rocks states that it is probably of Triassic age and that "it may be older, though this is regarded as highly improbable". From the evidence to be presented it will be seen that the Koipato formation is certainly older than Middle Triassic, and, according to recent studies by Wheeler (1937, p. 394; 1939, p. 107), is at least in part of Permian age.

At a number of places in the West Humboldt Range the Koipato effusive rocks are overlain disconformably, or at least without any perceptible angular discordance, by the bluish-black bituminous limestones of the Star Peak formation, and for example, in Congress Canyon, a tributary to the Unionville Canyon, these limestones near the base are richly fossiliferous, containing such distinctive cephalopod genera as *Cuccoceras, Balatonites,* and *Acrochordiceras.* This faunal assemblage, occurring below the *Ceratites trinodosus* beds, represents the lower part of the Anisic stage or the lowest division of the Middle Triassic (Muschelkalk). The volcanic rocks disconformably below this lower Anisic faunal horizon are therefore older than Middle Triassic. The pre-

Standard European section				Hawthorne and Tonopah quadrangles, Nevada	
System	Series	Stages	Faunal zones	Formations	Fossils
Jurassic.	Middle Jurassic.	Bajocian.			
	Lower Jurassic or Lias. — Upper Lias.	Toarcian.	*Lytoceras jurense.* *Hildoceras bifrons.* *Harpoceras falciferum.* *Dactylioceras tenuicostatum.*	Dunlap formation.	*Harpoceras* sp.
	Lower Jurassic or Lias. — Middle Lias.	Pliensbachian.	*Amaltheus spinatus.* *Amaltheus margaritatus.* *Prodactylioceras davoei.* *Tragophylloceras ibex.* *Uptonia jamesoni.* *"Deroceras" armatum.*	Disconformity (Locally conformity)	*Plicatostylus* cf. *P. gregarius.* *Enderoceras* ex. gr. *E. armatum.*
	Lower Jurassic or Lias. — Lower Lias.	Sinemurian.	*Echioceras raricostatus.* *Oxynoticeras oxynotus.* *Asteroceras obtusum.* *Arnioceras semicostatum.* *Agassiceras scipionianum.* *Ammonites bucklandi.*	Sunrise formation.	*Coroniceras* sp. *Ammonites bisulcatus.*
		Hettangian.	*Schlotheimia angulata.* *Psiloceras planorbis.*	Transition	*Exphyllites struckmanni,* *Psiloceras psilonotum.*
Triassic.	Upper Triassic.	Rhaetic.	*Pteria contorta* or *Choristoceras marshi.*		*Choristoceras marshi.*
		Noric.	*Sirenites argonauta.* *Pinacoceras metternichi.* *Cyrtopleurites bicrenatus.* *Cladiscites ruber.* *Sagenites giebeli.* *Discophyllites patera.*	Gabbs formation.	*Arcestes gigantogaleatus.* *Pinacoceras metternichi.* *Sagenites giebeli.*
		Karnic.	*Tropites subbullatus.* — *Juvavites* subzone. — *Trachyceras* subzone. *Carnites floridus.* *Trachyceras aonoides.* *Trachyceras aon.*	Conformity Luning formation.	*Gümbelites-Styrites.* *Tropites.* *Carnites*—*"Cardita."* Coral reefs.
	Middle Triassic.	Ladinic.	*Protrachyceras archelaus.* *Protrachyceras reitzi.*	Grantsville formation.	*Ceratites* sp. *Pecten* cf. *P. alberti.* *Hoernesia* aff. *H. socialis.*
		Anisic.	*Ceratites trinodosus.* *Ceratites binodosus.* *Parapopanoceras.*	Excelsior formation.	*Pecten* aff. *P. tirolicus.* *Hoernesia* aff. *H. socialis.*
	Lower Triassic.	Scythic.	*"Pseudharpoceras!"* *Columbites.* *Tirolites.* *Meekoceras.* — *Anasibirites* subzone. — *Owenites* subzone. — *Pseudosageceras* subzone. *Genodiscus.* *Otoceras.*	Contact not exposed Candelaria formation.	*Proptychites.* *Claraia clarai.*
Permian.					

ndard European section		Hawthorne and Tonop Nevada	
Stages	Faunal zones	Formations	
cian.		?---?---?	
cian.	*Lytoceras jurense.* *Hildoceras bifrons.* *Harpoceras falciferum.* *Dactylioceras tenuicostatum.*		
sbachian.	*Amaltheus spinatus.* *Amaltheus margaritatus.* *Prodactylioceras davoei.* *Tragophylloceras ibex.* *Uptonia jamesoni.* *"Deroceras" armatum.*	Dunlap formation. Disconformity (Locally conformity)	
murian.	*Echioceras raricostatum.* *Oxynoticeras oxynotum.* *Asteroceras obtusum.* *Arnioceras semicostatum.* *Agassiceras scipionianum.* *Ammonites bucklandi.*	Sunrise formation.	
tangian.	*Schlotheimia angulata.* *Psiloceras planorbis.*	Transition	
etic.	*Pteria contorta* or *Choristoceras marshi.*		
ic.	*Sirenites argonauta.* *Pinacoceras metternichi.* *Cyrtopleurites bicrenatus.* *Cladiscites ruber.* *Sageceras gicbeli.* *Discophyllites patens.*	Gabbs formation.	
		Conformity	
nic.	*Tropites subbullatus.* — *Juvavites* subzone. / *Trachyceras* subzone. *Caruites floridus.* *Trachyceras aonoides.* *Trachyceras aon.*	Luning formation.	
inic.	*Protrachyceras archelaus.* *Protrachyceras reitzi.*	Grantsville formation.	
nic.	*Ceratites trinodosus.* *Ceratites binodosus.* *Parapopanoceras.*	Excelsior formation.	
		Contact not exposed	
thic.	*"Pseudharpoceras?"* *Columbites.* *Tirolites.* *Meekoceras.* — *Anasibirites* subzone. / *Owenites* subzone. / *Pseudosageceras* subzone. *Genodiscus.* *Otoceras.*	Candelaria formation.	

TABLE 3.—*Correlation table*

rne and Tonopah quadrangles, Nevada		West Humboldt Range, Nevada		Shasta County, California		Inyo Range, California		Southern and	
ations	Fossils	Formations	Fossils	Formations.	Fossils	Formations	Fossils	Formatio	
				——?——	?			Nugget sand	
? ---- ?								?---?	
	Harpoceras sp.			Potem formation.					
ormation.	*Plicatostylus* cf. *P. gregarius.*								
formity				?—?—?					
onformity)	*Enderoceras* ex. gr. *E. armatum.*								
		——?——	?	Modin formation.	*Pholadomya nevadana, Entolium meeki*, etc.				
ormation.	*Coroniceras* sp. *Ammonites bisulcatus.*	Limestone of Muttleberry Canyon.	*Arnioceras humboldti.*						
	Euphyllites struckmanni, Psiloceras psilonotum.								
sition		Probable conformity (see text).		Unconformity (?)		?	?		
	Choristoceras marshi.								
ormation.	*Arcestes gigantogaleatus, Pinacoceras metternichi.*		*"Pseudomonotis" subcircularis.*	Brock shale.	*"Pseudomonotis" subcircularis.*		No fossils.		
				Conformity		Volcanic rocks. (Unnamed.)	(May be of Middle, Upper, or post-Triassic age.)		
	Sagenites giebeli.				*Spiriferina.* Coral reef.			?—	
ormity	*Gümbelites-Styrites.*	Star Peak formation.		Hosselkus limestone.					
	Tropites.		*Tropites.*		*Tropites.*				
formation.	*Carnites*—"*Cardita.*" Coral reefs.		Coral reef.	Conformity	*Halobia superba.* *Halobia rugosa.*				Ankare formatio
						?	?—?		
Grantsville formation.	*Ceratites* sp. *Pecten* cf. *P. alberti.* *Hoernesia* aff. *H. socialis.*		*Ceratites trinodosus.*	Pit shale.	*Ceratites* (?)		Whether or not entire Middle Triassic is represented by these beds is uncertain.		
?							*Parapopanoceras.*		
ot exposed	*Pecten* aff. *P. tirolicus.* *Hoernesia* aff. *H. socialis.*	Disconformity	*Acrochordiceras, Cuccoceras*, and *Beiatonites.*	Conformity		Limestones and shales. (Unnamed.)	*Acrochordiceras*	?	
formation.		Koipato formation.	(Volcanic rocks, in part or all (?) Permian.)	Dekkas andesite.	(No fossils.)		*Owenites.*		Thaynes gr
	Proptuchites. *Claraia clarai.*						*Pseudosageceras.*	?	
				——?	?				Woodside s
			Helicoprion.						

TABLE 3.—*Correlation table*

	Shasta County, California		Inyo Range, California		Southern Idaho, northern Utah, and western Wyoming		Arizona, Utah, and eastern Nevada		Parso
	Formations.	Fossils	Formations	Fossils	Formations	Fossils	Formations	Fossils	Form
	?	?			Nugget sandstone.	No fossils. Exact age not known.	?	?	
	Potem formation.				? ? ?	?			
	? ? ?						?	?	
	Modin formation.	*Pholadomya nevadana*, *Entolium meeki*, etc.					Glen Canyon group. (Wingate sandstone.)	(No fossils.)	Harb form
	Unconformity (?)		?	?			?		Disconfo
	Brock shale.	"*Pseudomonotis*" *subcircularis*.		No fossils.				(Continental red beds.)	Parso form
	Conformity		Volcanic rocks. (Unnamed.)	(May be of Middle, Upper, or post-Triassic age.)	?	?	Chinle formation.	Vertebrates; fresh-water mollusks.	600 feet conc
	Homelkus limestone.	*Spiriferina*. Coral reef. *Tropites*.				*Spiriferina* sp.			Valdez f
	Conformity	*Halobia superba*. *Halobia regens*.			Ankareh formation.		Conformity[1]		Base not
	Pit shale.	*Ceratites* (?)	?	Whether or not entire Middle Triassic is represented by these beds is uncertain.			? (Locally gradation.) Shinarump conglomerate.	(No fossils.)	
	Conformity		Limestones and shales. (Unnamed.)	*Parapopanoceras*. *Acrochordiceras*	?	*Columbites*. *Tirolites*.	Unconformity		
	Dekkas andesite.	(No fossils.)		*Owenites*. *Pseudosageceras*.	Thaynes group.	*Anasibirites*. *Owenites*. *Pseudosageceras multilobatum*. (*Meekoceras*.)	Moenkopi formation.	Continental deposits; lenses of marine sediments with *Meekoceras* (?).	
	?	?			? Woodside shale.	Pelecypods and *Lingula*.			

[1] Personal communication from J. B. Reeside.

	Parson Bay, British Columbia, Canada		Western Peace River, British Columbia, Canada		Alaska	
	Formations	Fossils	Formations	Fossils	Formations	Fossils
	Harbledown formation.	*Melanhippites.* *Arniotites.* *Arnioceratine ?*	Lower Bullhead Mountain formation.			
	Disconformity (†)		Contact not exposed		† †	(Possibly in part Jurassic.)
eds.)	Parson Bay formation.	"*Pseudomonotis*" *subcircularis.* *Maquoia alaskana.* *Juvavites* cf. *J. knowltoni.*		"*Pseudomonotis*" *subcircularis.*	McCarthy formation.	"*Pseudomonotis*" *subcircularis.*
	600 feet of section concealed.				†	
usks.		*Isastraea* cf. *I. vancouverensis.*		*Juvavites-Gricebachites.*	Unnamed rocks.	Coral reefs.
	Valdez formation.	(Mostly volcanic rocks.)			† Chitistone limestone.	*Tropites.*
	Base not exposed.	†	Schooler Creek formation.		† Unnamed rocks.	"*Dawsonites.*"
				Lobites.	† Unnamed rocks.	*Daonella.*
ites;					† Nikolai greenstone.	Greenstones and lavas. No fossils. (Possibly all or only in part Permian.)

Middle Triassic age of the Koipato rocks is now confirmed through Wheeler's (1937, p. 394; 1939, p. 107) discovery of *Helicoprion* in the Rochester trachyte about 7000 feet below the Middle Triassic limestones. *Helicoprion* is diagnostic of the middle part of the lower Permian (Artinskian of the European section) so that at least the lower part of the Koipato is of late Paleozoic age. It is hoped that further work in the West Humboldt Range will throw more light on the age determination of the upper part of the Koipato.

In the Inyo Mountains, about 100 miles south of the Hawthorne and Tonopah quadrangles, Knopf and Kirk (Knopf, 1918, p. 48) describe a thick section of volcanic rocks above the Middle Triassic limestones and class them as Upper Triassic. They admit, however, that these rocks may be of Middle, Upper, and even post-Triassic age, which makes correlation of that section with the Excelsior formation of the Hawthorne and Tonopah quadrangles quite uncertain.

In the Shasta region of California, there is, according to Diller (1906), over 1500 feet of volcanic rocks (Dekkas andesite). These lie unconformably upon the Permian Nosoni formation and are in turn overlain by the Middle and Upper Triassic Pit shale. Hinds (1932, p. 404) reports that in the upper part of the Nosoni, immediately below the contact with the Dekkas andesite, lavas and bedded tuffs are interstratified with the marine sedimentary rocks. The Dekkas andesite is generally regarded as Triassic. Since, however, the Nosoni formation is not of the latest Permian age, the lower stratigraphic limit of the overlying volcanic rocks becomes rather indefinite, and the Dekkas andesite may be, at least in part, of late Permian age.

Crickmay (1928) reported over 5000 feet of volcanic rocks from the Parson Bay area in British Columbia which he correlated with the Valdez formation (? = Valdes group of the Canadian Geological Survey). On the basis of a coral referred by Crickmay to *Isastraea vancouverensis* these volcanic rocks were dated by him as Noric in age.

It has already been shown (Muller, 1936) that the Triassic corals in western North America, formerly correlated with the Noric "interregional zone", have a much wider stratigraphic range and may be of Karnic and even earlier age; it is therefore possible that the British Columbia pyroclastic rocks may be in part equivalent to the Excelsior formation of Nevada, but there is as yet no evidence for this correlation.

Metamorphosed volcanic rocks associated with Triassic sedimentary rocks are also widely distributed in Alaska (Martin, 1926), but unfortunately their stratigraphy is as yet inadequately known.

It is therefore evident that much field work still remains to be done

before it will be possible to establish the exact relations between these different volcanic sections.

GRANTSVILLE FORMATION (MIDDLE TRIASSIC)

Definition and areal extent.—Since the publication of the preliminary account of the Triassic and Jurassic formations of the area (Muller and Ferguson, 1936), a new cartographic unit—the Grantsville formation—has been added to the sequence of Triassic rocks in the Hawthorne and Tonopah quadrangles.

The Grantsville formation is named after a mining camp on the west flank of Shoshone Mountains in the north-central part of the Tonopah quadrangle. The type locality of the formation is on the crest of the spur separating the mouth of Union Canyon from the next large canyon to the north.

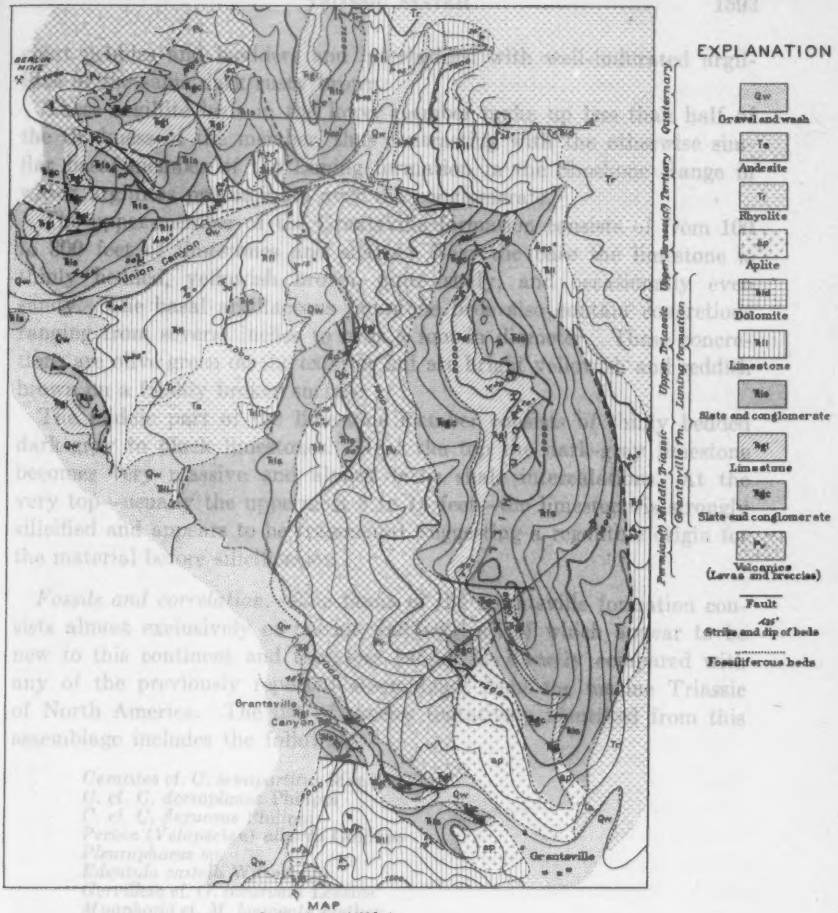
In addition to the type locality the Grantsville formation outcrops also on the ridge due north of Grantsville, at the mouth of Grantsville Canyon, and on a low isolated hill just south of Union Canyon (Pl. 2).

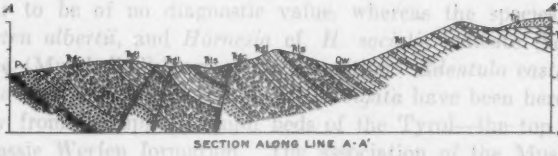
Two members are recognized: a lower member of conglomerate with interbedded argillite and an upper member of limestone with minor shales.

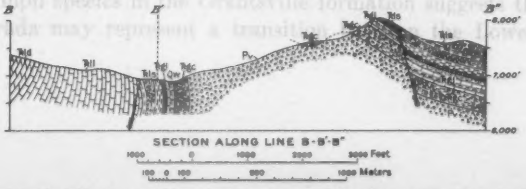
The Grantsville formation has not been recognized in any other part of the two quadrangles, with the possible exception of a small area in the northern part of the Paradise Range. It is likely, however, that further study of the Grantsville fauna may prove it to be a correlative of the lenticular marine sediments in the Excelsior formation of the Gillis Range. This would imply an abrupt change from the thick section of volcanic rocks constituting the Excelsior formation in the Hawthorne quadrangle to the very thin and exclusively sedimentary Grantsville formation.

The Grantsville formation rests unconformably upon a thick section of altered volcanic rocks which have been tentatively placed in the Permian because of their proximity to the rocks of almost identical lithology in the Toyabe Range which are associated with sedimentary rocks of Permian age. The Grantsville formation is overlain unconformably by the Luning formation with only minor angular discordance.

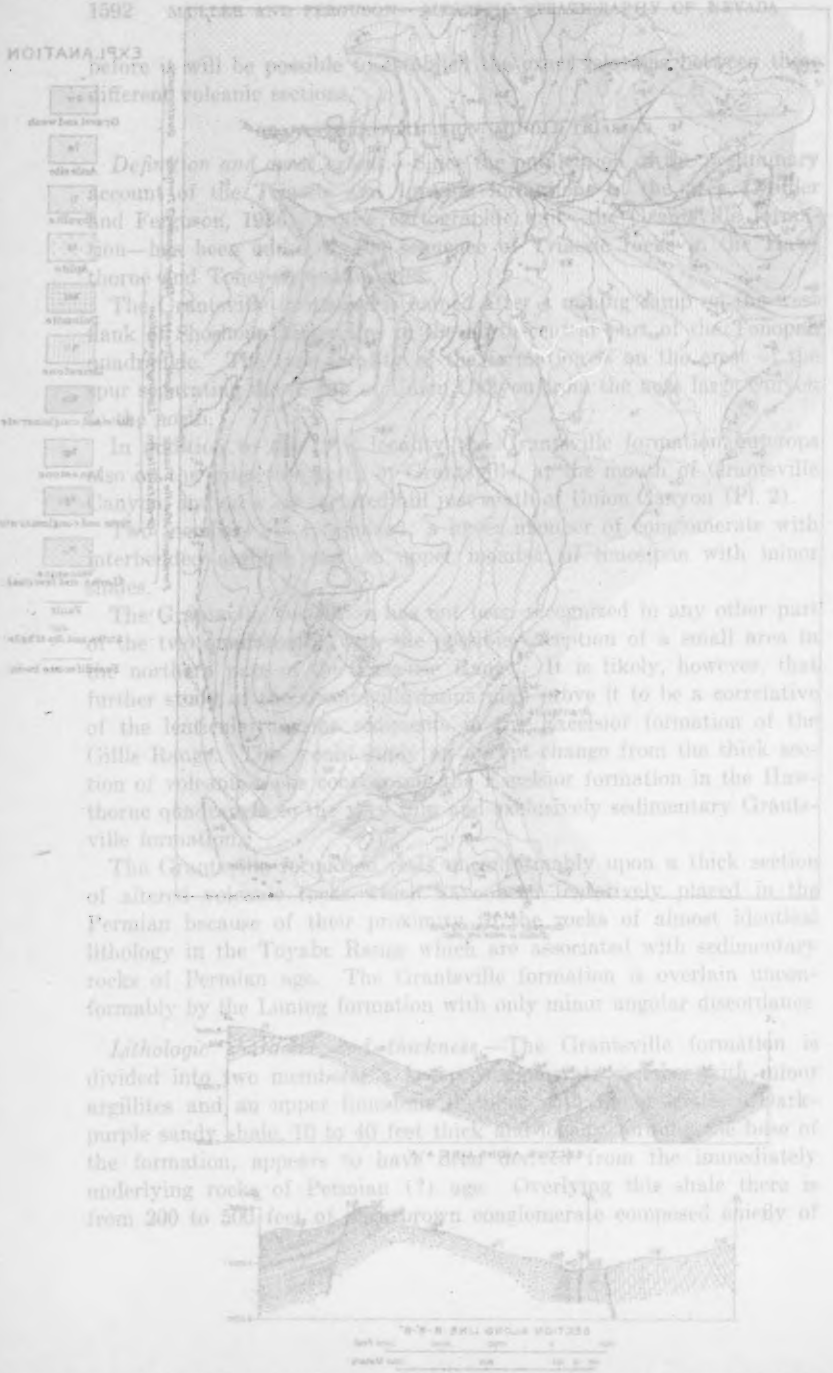
Lithologic character and thickness.—The Grantsville formation is divided into two members: a lower conglomerate member with minor argillites and an upper limestone member with minor shales. Dark-purple sandy shale, 10 to 40 feet thick and locally forming the base of the formation, appears to have been derived from the immediately underlying rocks of Permian (?) age. Overlying this shale there is from 200 to 500 feet of dark-brown conglomerate composed chiefly of







GEOLOGIC MAP OF PART OF SHOSHONE MOUNTAINS BETWEEN UNION CANYON AND GRANTSVILLE CANYON.



GEOLOGIC MAP OF PART OF SHOSHONE MOUNTAINS BETWEEN UNION CANYON AND GRANTSVILLE CANYON.

chert pebbles and boulders and interbedded with well-indurated argillites of the same dark rusty brown.

These argillite beds in the lower member make up less than half of the thickness of the member, thus contrasting with the otherwise similar basal member of the Luning formation in the Shoshone Range in which argillites predominate over the conglomerate.

The upper member of the Grantsville formation consists of from 100 to 300 feet of limestones and shales. Near the base the limestone is thinly bedded, yellowish brown, quite shaly, and occasionally even sandy. The basal argillaceous limestone beds also contain concretions ranging from several inches to over a foot in diameter. These concretions are olive green on the exterior but are bright yellowish and reddish brown on a freshly broken surface.

The middle part of the limestone member consists of thinly bedded dark-gray to black limestones. Near the top the dark-gray limestone becomes very massive and almost lacks shale intercalations. At the very top—usually the uppermost 5 to 15 feet—the limestone is strongly silicified and appears to be fragmental, suggesting a regolithic origin for the material before silicification.

Fossils and correlation.—The fauna of the Grantsville formation consists almost exclusively of marine mollusks, all of which appear to be new to this continent and therefore cannot be directly compared with any of the previously reported assemblages from the marine Triassic of North America. The list of species tentatively identified from this assemblage includes the following:

Ceratites cf. *C. semipartitus* Montfort
C. cf. *C. dorsoplanus* Philippi
C. cf. *C. flexuosus* Philippi
Pecten (*Velopecten*) *albertii* Goldfuss
Pleurophorus sp.
Edentula castelli Wittenburg
Gervilleia cf. *G. incurvata* Lepsius
Myophoria cf. *M. laevigata* Ziethen
Hörnesia cf. *H. socialis* (Schlotheim)
Spiriferina sp.

Of this list, *Pleurophorus* and *Spiriferina*, specifically unidentifiable, appear to be of no diagnostic value, whereas the species of *Ceratites*, *Pecten albertii*, and *Hörnesia* cf. *H. socialis* indicate the Middle Triassic (Muschelkalk) age of the formation. *Edentula castelli*, *Gervilleia incurvata*, and *Myophoria* cf. *M. laevigata* have been heretofore known only from the upper Campil beds of the Tyrol—the top of the Lower Triassic Werfen formation. The association of the Muschelkalk with the Campil species in the Grantsville formation suggests that these beds in Nevada may represent a transition between the Lower and Middle

Triassic. Such a conclusion, however, would seem to be rather premature, and in view of the fact that the Muschelkalk species appear to predominate over the Campil species it is preferable, at least for the present, to regard the Grantsville fauna as of early Middle Triassic age.[2]

We have thus arrived indirectly at the conclusion that the Grantsville fauna is approximately contemporaneous with the Excelsior assemblage which also contains a mixture of Middle and Lower Triassic forms. The two faunas are also similar in that they both show certain affinities with the "Germanic" development of the Triassic. Only one species—*Hörnesia* cf. *H. socialis*—is common to these two faunas. The lack of closer ties between the two assemblages in this area is due in large part to the inadequacy of the collections. This is particularly true of the Excelsior formation, from which was obtained only a small collection, in which the fossils are much distorted and hard to separate from the matrix. The lack of closer similarity between the two faunas may also be due to differences in facies or perhaps even to a slight difference in age.

The Grantsville formation is considered equivalent to the lower Muschelkalk and may be thus indirectly correlated with the lower part of the Star Peak formation in the West Humboldt Range (Table 3). But, as is also the case with the Excelsior formation, the differences in the faunal facies of the two units preclude accurate matching of the sections.

LUNING FORMATION (UPPER TRIASSIC)

Definition.—The Luning formation derives its name from a small settlement in Soda Spring Valley in the central part of the Hawthorne quadrangle. The rocks included in this formation are dominantly dark dolomites and limestones with a subordinate amount of interbedded shales, argillites, and coarser clastic rocks. In general the formation shows a marked range in lithology laterally as well as vertically. It rests unconformably upon the Excelsior formation of Middle Triassic age and in the Shoshone Mountains it overlies the Middle Triassic Grantsville formation with an erosional unconformity but with only very slight angular discordance. The upper limit of the Luning formation in most places is marked by a sharp change from dark massive dolomites and limestones to the conformably overlying thinly bedded shales and limestones of the Gabbs formation. Locally the Luning formation is overlain with marked angular discordance by the Lower Jurassic Dunlap formation.

As the Gabbs formation is considered equivalent to the Noric and

[2] In the Recoaro district of the Alps, the basal Muschelkalk limestones, corresponding to the *Dadocrinus gracilis* limestones, likewise contain mollusks which are closely allied to the Lower Triassic species of the Campil beds (Diener, 1912, p. 77).

Rhaetic stages of the European section, the top of the Luning formation is correlated with the top of the Karnic stage. Locally where this horizon is not marked by a change in lithology and the fossils are few and not diagnostic, as, for example, in the Shoshone Mountains and in the central part of the Cedar Mountains, it is not possible to map the Karnic-Noric boundary; beds corresponding in age to the Noric and possibly even to the Rhaetic of the European section are also included in the Luning formation.

The total thickness of the formation may approach 10,000 feet, although the thickest single section (Pilot Mountains) is estimated at approximately 8000 feet.

Lithologic character.—The lithology and thickness of the Luning formation are somewhat varied from range to range and even in different parts of the same range. In general, gray to black limestones and dolomites with subordinate argillite are the dominant rocks, but the proportion of one type of rock to the other may change very markedly within a few miles. In some areas, such as the overthrust mass in the Paradise Range, the formation is composed almost entirely of massive, dark-gray dolomite. On the other hand, argillites with interbedded conglomerates interpreted as near-shore sediments form a prominent feature of the formation in other areas, such as the Pilot Mountains.

Between the belt of these near-shore sedimentary rocks and the dolomites, which suggest an off-shore facies, there are sediments of intermediate facies consisting of argillite and limestone. These include exposures of Luning northeast of Luning, the Luning limestones of the Gillis Range, and parts of the Luning section in the Shoshone and Cedar Mountains and in the Garfield Hills. Scattered occurrences of intercalated coral reefs over the entire area suggest general shallowness of the sea with some irregularities of the strand line—perhaps an archipelago-like condition. It seems reasonable to conclude that the southern and, in part, eastern limits of the Luning outcrops (Pl. 1) roughly mark the shore of the sea in which they were deposited, with a land mass situated to the south and, less definitely, to the east. The area in which the Luning sediments were deposited was dominantly sinking in Luning time, but locally there were marked oscillations evidenced by a frequent recurrence of conglomerates in the near-shore belt (Pilot Mountains), by a repetition of banks of *Alectryonia montis-caprilis,* and by a repeated occurrence of thin coral reefs.

From the foregoing it may be seen that the Luning formation is very heterogeneous in constitution, but this in itself determines its unity. In areas where owing to structural breaks the Luning formation is incom-

plete, and is represented by only one or two types of rock, its identity can be, as a rule, readily established on the basis of fossils it contains as well as from the structural relations to the adjacent formations.

The lithology of the Luning formation at the principal localities may be summarized as follows:

Pilot Mountains: In the central part of the Pilot Mountains, the type locality of the formation, the Luning has a total thickness of approximately 8000 feet. The formation includes three members (Pl. 3).

The upper member consists essentially of limestone with subordinate dolomite and minor interbedded slate and argillite. The top is concealed beneath the Tertiary volcanic rocks, but the exposed thickness is approximately 2500 feet.

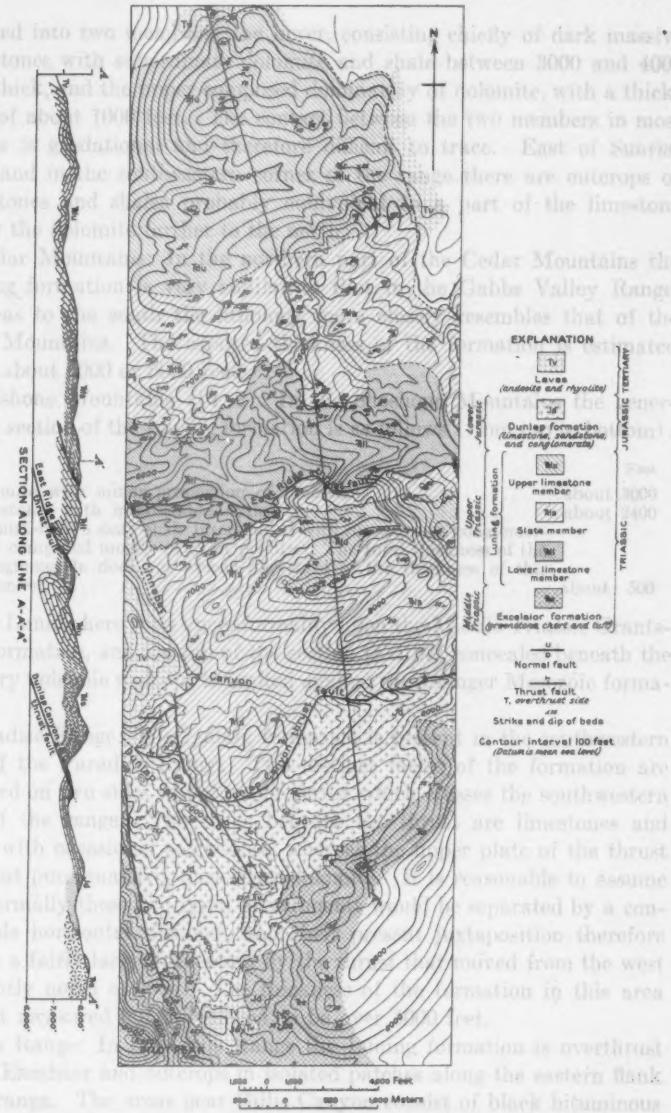
The middle member, with a maximum thickness of about 3000 feet, consists of argillites, slates, and conglomerates made up exclusively of chert pebbles derived from the Excelsior to the south. The proportion of conglomerate increases southward, and near the head of Cinnabar Canyon conglomerate forms almost half of the middle member.

The lower member consists essentially of rather thin-bedded limestone but near the top contains interbedded slates and several coral reefs (bioherms). As there is no section of the lower member unbroken by thrusts, the thickness is uncertain but is estimated at about 2500 feet.

The base of the Luning is exposed only in the complexly faulted area in the northwestern part of the range. Within one of the thrust plates a basal conglomerate of chert pebbles rests with angular unconformity on the chert of the Excelsior formation and grades upward into argillite similar to that of the middle member. Within other thrust plates massive greenstone of the Excelsior is directly overlain by the lower limestone member of the Luning formation without appreciable basal conglomerate.

Garfield Hills: In the Garfield Hills two slate members are recognized in the Luning formation, with a limestone member between them. Locally the base of the formation is marked by a conglomerate. Similar conglomerate also occurs sporadically throughout the slate members. To the west in the thrust blocks of the Luning formation the slate members appear to be much thinner than in the eastern part of the Garfield Hills. To the southwest a lava flow altered to greenstone is present within the Luning formation, a feature which has not been observed in the Luning formation elsewhere. As no complete section of the Luning formation is present in the Garfield Hills its total thickness here cannot be measured, but the exposed thickness does not exceed 4000 or 5000 feet, which is about half that of the Pilot Mountains.

Gabbs Valley Range: In the Gabbs Valley Range the Luning formation attains a maximum thickness of approximately 5000 feet and may be



GEOLOGIC MAP OF NORTH-CENTRAL PART OF PILOT MOUNTAINS.

divided into two members: the lower, consisting chiefly of dark massive limestones with subordinate dolomite and shale between 3000 and 4000 feet thick, and the upper composed dominantly of dolomite, with a thickness of about 1000 feet. The contact between the two members in most places is gradational and therefore difficult to trace. East of Sunrise Flat and in the southeastern corner of the range there are outcrops of limestones and shales probably equivalent to a part of the limestone below the dolomite farther to the north.

Cedar Mountains: In the northern part of the Cedar Mountains the Luning formation is very similar to that in the Gabbs Valley Range, whereas to the south the lithology more closely resembles that of the Pilot Mountains. The exposed thickness of the formation is estimated to be about 4000 or 5000 feet.

Shoshone Mountains (Pl. 2): In the Shoshone Mountains the generalized section of the Luning formation is as follows (from top to bottom):

	Feet
Dolomite with minor proportion of limestone	about 3000
Limestones with interbedded shales	about 2400
Greenish-brown slaty shale interbedded with rusty-brown conglomerate composed mostly of chert pebbles. The total thickness of the conglomerate does not exceed one third of the thickness of this member	about 500

The Luning here rests unconformably upon the Middle Triassic Grantsville formation, and the top of the section is either concealed beneath the Tertiary volcanic rocks or is faulted against the younger Mesozoic formations.

Paradise Range: The Luning formation is present in the southwestern part of the Paradise Range. Two distinct facies of the formation are observed on two sides of the major thrust which crosses the southwestern part of the range. The rocks beneath the thrust are limestones and shales with occasional sandstones, whereas the upper plate of the thrust is almost pure, massive, dark-gray dolomite. It is reasonable to assume that normally these two types of sediments would be separated by a considerable horizontal distance, and their present juxtaposition therefore implies a fairly large dislocation by the thrust that moved from the west or slightly north of west. The thickness of the formation in this area was not measured but is believed to be over 5000 feet.

Gillis Range: In the Gillis Range the Luning formation is overthrust by the Excelsior and outcrops in isolated patches along the eastern flank of the range. The areas near Gillis Canyon consist of black bituminous limestones and shales that resemble closely the Luning rocks in the Gabbs Valley Range (northeast of Luning), the northern part of Cedar Mountain, and also some of the strata in the lower part of the formation in

the Shoshone Mountains. Only a small part of the total thickness of the formation is exposed in the Gillis Range.

Toyabe Range: At the south tip of the Toyabe Range several hundred feet of recrystallized limestones with interbedded slate were mapped as the Luning formation solely on the basis of general lithologic similarity to other fossiliferous Luning sections (Fig. 2).

Fossils and age.—Large collections of marine invertebrates were obtained from several horizons in the Luning formation. At least three different faunal facies are represented by the material: a near-shore pelecypod facies, a coral-reef facies, and an off-shore ammonite facies.

Of the pelecypod faunas in the Luning formation the *Alectryonia montis-caprilis* fauna is the most widespread and is also of considerable stratigraphic range. In the Pilot Mountains it occurs sporadically through more than half of the Luning formation and it is recognized in the Shoshone Mountains, Gabbs Valley Range, Garfield Hills, Cedar Mountains, and Paradise Range. This fluted oyster is usually associated with *Myophoria* cf. *M. kefersteini* (Münster), *M.* cf. *M. whateleyae* Buch, and occasionally with *"Cardita"* sp.[3] As a rule the remains of these pelecypods occur in regular shell banks but have been intensely compressed and distorted by the folding of the containing rocks, and the original shell material is seldom preserved. It is doubtful if *Alectryonia montis-caprilis* is a very reliable stratigraphic marker, although to date its occurrence has been limited to beds equivalent to the upper Karnic of Europe. The two species of *Myophoria* are restricted to strata equivalent to the middle Karnic. Thus the age of the entire assemblage from Nevada may correspond to middle and upper Karnic.

Other pelecypods occur sparingly throughout the Luning formation, and their remains are usually poorly preserved, so that for many not even generic identification is possible. The genus *Halobia,* a common Upper Triassic bivalve, is recognized in the Shoshone Mountains and in the Cedar Mountains. Another pelecypod worthy of special note is the thick-shelled *Trichites* (=*"Pinnigena")* which until publication of the recent article by Stanton (1926, p. 485) had been known only in the Jurassic and Cretaceous. Several beds of impure limestone overlying the coral reefs and interstratified with them contain a large number of these shells with their characteristic prismatic structure preserved. In shape and sculpture the Nevada species resembles somewhat the Jurassic *T. seebachi* (Böhm).

Reef-building corals were collected from the Luning formation in the

[3] The study of the hinge of this bivalve revealed some differences from that of the true *Cardita.* It is also likely that some of the Alpine forms generally assigned to *Cardita* may belong to another genus.

Pilot Mountains, the Gillis Range, the Garfield Hills, the Shoshone Mountains, and the Cedar Mountains. Of these by far the best collection was obtained from the Pilot Mountains, where the corals occur at several stratigraphic levels, all within about 200 feet. The conclusions reached from the study of these corals have already been presented (Muller, 1936, p. 202-208), and the subject will be therefore only briefly reviewed at this time.

Corals from the Pilot Mountains were first collected by Turner in about 1900 and were originally dated by Smith as Jurassic (Spurr, 1903, p. 104). Smith (1912, p. 95) modified his conclusion regarding the age of these corals and correlated them with the lower Noric of the European Upper Triassic. Later, he (Smith, 1927) enlarged upon the list of corals from the Triassic of North America, still correlating them with the "Interregional Coral zone of the Lower Noric". As a result of recent work on the Triassic rocks of the Pilot Mountains more than a dozen species of corals are now recognized from the Dunlap Canyon locality. The study of this new material reveals a very striking similarity between the Dunlap Canyon assemblage and the so-called Noric corals of Shasta County, California. The general resemblance between the two faunas is so strong that, at first glance, there appears to be no reason to doubt the Noric age of these Nevada corals. However, in spite of this marked similarity the Dunlap Canyon corals should be correlated with the Karnic. This determination is based on brachiopods and mollusks associated with the coral reefs and also found stratigraphically above the corals.

Among the brachiopods found in association with reef-building corals the following species have been recognized:

Spiriferina gregaria Suess
S. peneckei Bittner
Terebratula debilis Bittner
T. julica Bittner
T. suborbicularis Münster var. *typica* Bittner

These brachiopods are known to occur in Europe only in strata of medial and early Karnic age.

Ammonites, although absent from the coral reefs, are nevertheless represented by 16 fragments collected from the beds immediately above and below the coral beds. Two ammonites from a shale between the two coral beds are identified as *Carnites* and *Klamathites*, both of which are characteristic of the Karnic stage in Europe. Other fragments of ammonites collected from beds a few feet above the uppermost coral reef belong to *Juvavites (Anatomites)* which is also of Karnic age.

Perhaps the strongest evidence for the Karnic age of the Dunlap Canyon corals is furnished by the pelecypod assemblage obtained from the argillite member of the Luning formation (Pl. 3) and also from the

upper part of the upper limestone member at stratigraphic levels from 1000 to 4000 feet above the coral horizons.

The molluscan assemblage includes the following species:

Myophoria cf. *M. kefersteini* Münster
M. cf. *M. whateleyae* Buch
Alectryonia montis-caprilis Emmrich
"Cardita" sp.

The two species of *Myophoria* are, in the Alpine Triassic, restricted to the middle Karnic, whereas *Alectryonia montis-caprilis* is usually found in the upper but not the uppermost Karnic. The occurrence of these mollusks at a level so far above the coral horizons proves that the Dunlap Canyon coral reefs are either equivalent to the medial or possibly even the early Karnic.

It is therefore rather significant that the reef corals from Dunlap Canyon, in spite of their difference in age, show a remarkable resemblance to the Noric assemblage from the Shasta region, California. This resemblance is so striking that Smith, who had studied the Triassic corals of North America, and also examined some of the writers' material, firmly believed that the Nevada corals were of Noric age.

Recent studies of the older Triassic rocks in Nevada, particularly in Desatoya (New Pass) Range, show also that it is equally difficult to distinguish corals of middle Triassic age from those of Upper Triassic age.

In conclusion it may be said that: (1) The coral reefs near the top of the lower limestone member of the Luning formation are not Noric, as was held by Smith, but Karnic; (2) these Karnic reef-building corals from the Luning formation show a remarkable resemblance to the Noric corals of California as well as to the Middle Triassic corals of Desatoya (New Pass) Range; (3) the reef-building corals alone cannot be used for exact correlation of Triassic strata in widely separated areas; and finally (4) in view of the above conclusions, it is necessary to verify the stratigraphic relations of a number of coral horizons in the Triassic of western North America, which appears to have been correlated with the Noric on insufficient evidence.

The *Carnites* fauna was collected in Union Canyon, Shoshone Mountains, from brown impure limestone interbedded with shales about 300 feet above the conglomerate (Pl. 2). The most common forms in the assemblage are:

Juvavites (*Anatomites*) *böhmi* Gemmellaro
J. (*A.*) *formosus* Gemmellaro
J. (*A.*) *elegans* Gemmellaro
Klamathites cf. *K. kellyi* Smith
Arcestes sp.
Carnites sp.
"Cardium" arcaeformis Gabb (= *"Cardita"*?)

Myophoria cf. *M. decussata* Münster
Nucula sp.

This fauna is distinctive of the Karnic stage of the European Upper Triassic. It also shows some resemblance to the Karnic fauna of California as evidenced by the presence of *Klamathites* cf. *K. kellyi* Smith and *Juvavites (Anatomites) böhmi* Gemmellaro. The latter is almost conspecific with *J. (A.) subintermittens* Hyatt and Smith.

Overlying conformably the *Carnites*-bearing limestones and about 230 feet higher in the section is a second fossiliferous horizon from which the following species are identified:

Tropites cf. *T. subbullatus* Mojsisovics
T. cf. *T. welleri* Smith
T. cf. *T. ursensis* Smith
Anatropites sp.
Hoplotropites cf. *H. ornatissimus* (Gemmellaro)
Juvavites (*Anatomites*) *inflatus* Gemmellaro
Arcestes sp.
Klamathites sp.

This assemblage is characteristic of the late Karnic *Tropites subbullatus* zone which is an interregional zone of world-wide distribution. It has been previously recognized in North America (Alaska, British Columbia (?), Oregon, California, and Mexico); in California it was subdivided by Smith (1927, p. 5) into two subzones, the lower or *Trachyceras* subzone and the upper or *Juvavites* subzone. The new fauna from Nevada is more closely related to that of the upper or *Juvavites* subzone of Smith.

A representative *Tropites* fauna was also obtained from a locality in the Gabbs Valley Range, about 7 miles northeast of Luning. Abundant but rather poorly preserved fossils representing a small number of species came from a 20-foot thickness of black crystalline limestone interbedded with thin seams of jet black shale. Because of the poor preservation specific identification was possible for only a few forms. Nevertheless the assemblage as a whole is very distinctive and diagnostic. From an horizon near the base of this 20-foot thickness were obtained several species of *Tropites* of the group of *Tropites welleri* Smith and *T. reticulatus* Smith, both of which, according to Smith (1927, p. 27), are restricted to the *Juvavites* subzone of the *Tropites subbullatus* zone and were heretofore known only from the Hosselkus limestone in Shasta County, California.

In the upper part of this 20-foot section were found several species of *Juvavites*, associated with *Cosmonautilus* cf. *C. pacificus* Smith. These too are restricted to the *Juvavites* subzone of the *Tropites subbullatus* zone.

Near Gillis (Wilhhorse) Canyon, Gillis Range, a small collection of ammonites was obtained from black bituminous limestones. Although no species of *Tropites* are present in this collection, the presence of such forms as *Juvavites kellyi* Smith, *Juvavites* cf. *J. damesi* Mojsisovics, *Discophyllites* ? sp., and *Cardita* cf. *C. pichleri* Bittner suggest a late Karnic age for the faunule, corresponding to the *Juvavites* subzone of the *Tropites subbullatus* zone. A similar assemblage was recognized in the Cedar Mountains about half a mile southwest of the Simon Mine.

The *Gümbelites* fauna occurs about 350 feet above the *Tropites subbullatus* zone. It is recognized in the Shoshone Mountains, the Cedar Mountains, and the Garfield Hills. The best collection was obtained from black bituminous limestones interbedded with shales in Union Canyon, Shoshone Mountains.

The following species are recognized in this assemblage:

Gümbelites philostrati Diener
G. jandianus Mojsisovics
Griesbachites sp.
Styrites vermetus (Dittmar)
S. signatus (Dittmar)
S. heberti Mojsisovics
S. cf. *S. malayicus* Welter
Palicites mojsisovicsi Gemmellaro
Mojsisovicsites crassecostatus Gemmellaro
M. obsoletus Gemmellaro
M. tenuicostatus Gemmellaro
Arcestes sp.

This very interesting assemblage occurs within a thickness of only a few feet (the bulk of the material in the Shoshone Mountains was obtained from a bed 2 feet thick); nevertheless it contains elements which in other parts of the world are known to occur in the Karnic, in the so-called Karnic-Noric transition fauna, and in the Noric.

The genus *Gümbelites*, for instance, is represented by two species that in India and Timor come from beds of Noric age. The genus *Griesbachites*, according to Diener, ranges from Karnic to lower Noric. Its earliest appearance is recorded by Mojsisovics from the lower part of the *Trachyceras aonoides* zone (*Trachyceras austriacum* subzone), which is lower Karnic. McLearn (1930, p. 15) reports *Griesbachites* from the Schooler Creek formation in Canada, from beds which are "Noric and, in part at least, lower Noric." It is of interest to note that in Canada the genus *Griesbachites* is associated with *Drepanites, Stikinoceras, Discotropites, Juvavites, Anatomites, Hannaoceras*, and others. Of these, *Stikinoceras* is also recognized in the *Gümbelites* fauna in Nevada, but its presence throws little light upon the geologic age of the Nevada assemblage, for, as McLearn states, the Canadian fossils may not be all of the same date.

The first three species of *Styrites* were previously recognized by Mojsisovics in the Karnic, in the beds above the *Trachyceras austriacum* subzone, whereas the species of *Palicites* and *Mojsisovicsites* come from the undifferentiated Neotrias of Sicily and are consequently of little value for accurate dating of rocks.

This peculiar association of seemingly heterochronous elements in the Nevada assemblage presents a rather puzzling problem, the solution of which calls for a thorough restudy of the stratigraphic relations and ranges of all the genera and species involved. Pending further investigation of this interesting assemblage it is thought best to consider it equivalent to the upper Karnic or lower Noric, possibly transitional.

Correlation.—As has been already pointed out in the discussion of faunas, the Luning formation is the equivalent of the Karnic stage of the Alpine Triassic section. Ammonite faunas permit direct correlation with the late Karnic of the standard European section, but the absence of ammonites from the lower part of the Luning formation makes it difficult to establish accurately its lower stratigraphic limit in terms of ammonite zones of the European section. The occurrence of such pelecypods as *Myophoria decussata, M.* cf. *M. kefersteini,* and *M.* cf. *M. whateleyae* some distance above the base strongly suggests that probably beds equivalent to most of the lower part of the Karnic are also represented in the Luning formation.

In North America the Luning formation can be correlated with the Hosselkus limestone of Shasta and Plumas counties, California, and also with the middle part of the Star Peak formation of the West Humboldt Range, northwestern Nevada—the part which is delimited by the *Daonella dubia* zone below and by the *"Pseudomonotis"*-bearing beds above (Table 3).

Sedimentary rocks corresponding to the Luning formation are also known in central Oregon (Schenk, 1935, p. 401), British Columbia (McLearn, 1930, p. 3) (part of the Schooler Creek formation and possibly also on the Coast), and in Sonora, Mexico (Burckhardt, 1905).

Although there is no direct means to establish correlation, the Luning may also be the time equivalent of a part of the continental Chinle formation in Utah and Arizona which is of late Triassic age. The same applies to other continental deposits of supposed late Triassic age in the Rocky Mountains region.

GABBS FORMATION (UPPER TRIASSIC)

Definition.—The Gabbs formation derives its name from the Gabbs Valley Range where it is well exposed at several localities, particularly

along the front of the range, in the vicinity of Volcano Peak (Pl. 4; Pl. 5, fig. 1). The formation is typically developed and richly fossiliferous at the type locality in New York Canyon about 2000 feet southeast of Volcano Peak. Here the Gabbs formation is about 420 feet thick and is made up of shales and limestones which rest conformably upon the Luning formation and grade upward into the Sunrise formation of early Jurassic age.

The lower boundary of the Gabbs formation is clearly defined by a marked lithologic change from massive dolomites of the underlying Luning formation to thinly bedded shales with subordinate limestone. The upper limit of the formation, however, is less distinct. The uppermost beds of the Gabbs formation and the lowermost beds of the Jurassic sunrise formation are lithologically very similar and can be differentiated only on the basis of marine invertebrate faunas.

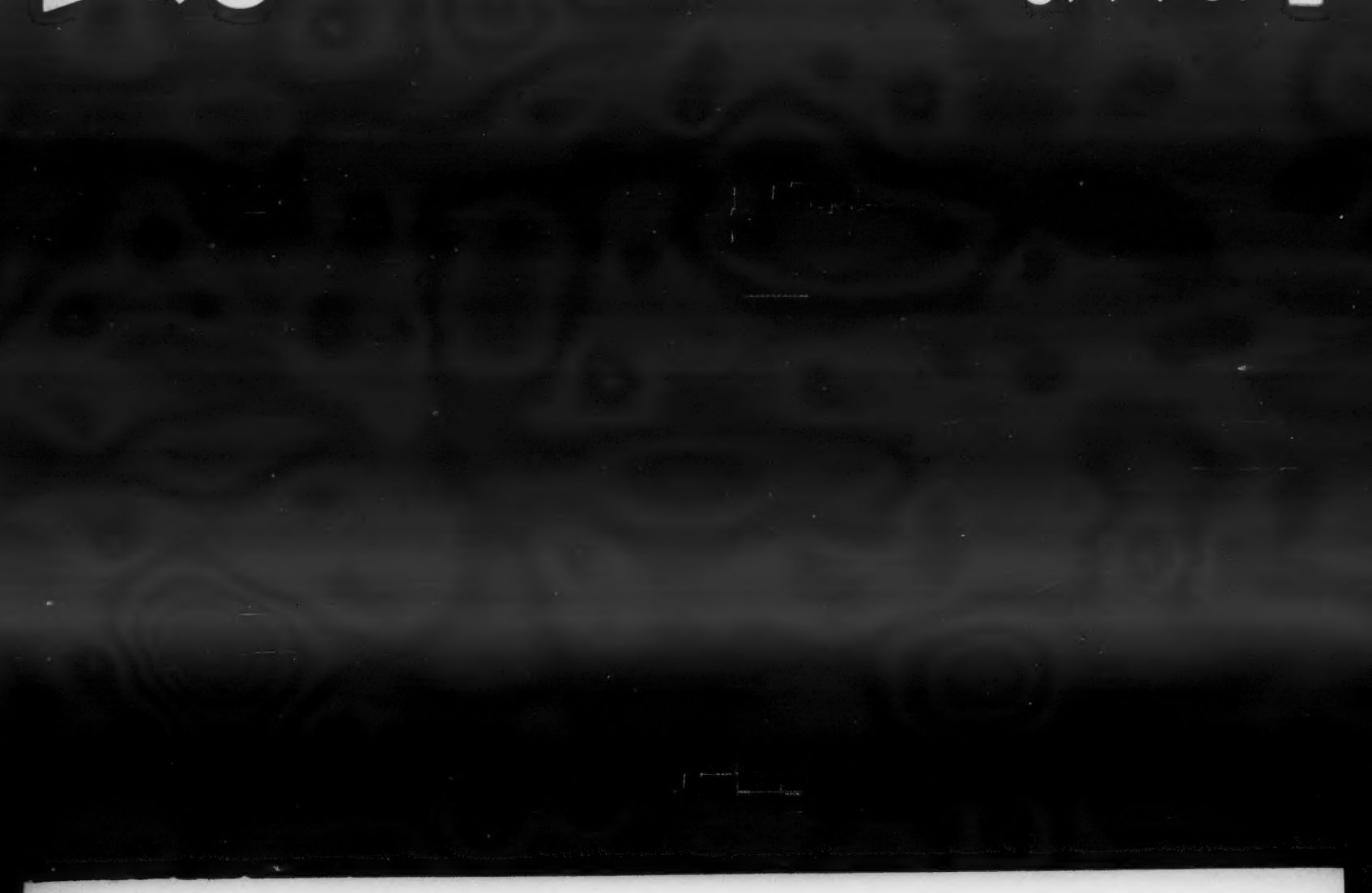
Other occurrences of the Gabbs formation are in the Garfield Hills, in the Paradise Range, and in the southwestern part of the Shoshone Mountains.

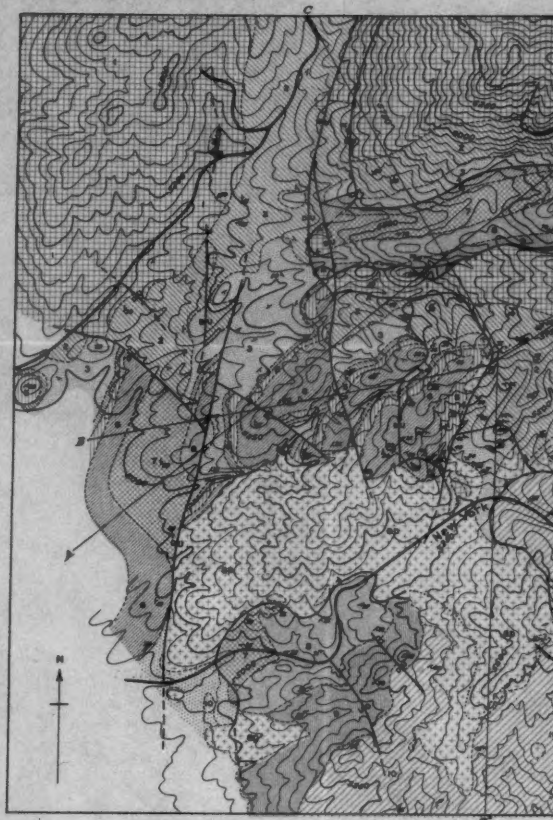
Lithologic character.—The Gabbs formation at the type locality in New York Canyon (Pl. 4) has been subdivided into three members. The lower member, about 290 feet thick, is composed of dark thinly bedded carbonaceous shale intercalated with beds of black impure limestone ranging in thickness from 4 inches to 2 feet. The shale and limestone in fresh exposures are dark gray to jet black but weather to a peculiar grayish purple which contrasts so strikingly with the color of the adjacent stratigraphic units as to render the lower member easily distinguishable even from a distance. One should not, however, place too much reliance upon this superficial character, for similar composition and effects of weathering are also encountered in the overlying Sunrise formation.

The middle member of the Gabbs formation, in contrast to the lower member, is composed predominantly of limestone, which is commonly brown and is shaly and even sandy. The individual layers of limestone range from 6 inches to 2 feet in thickness and are interstratified with thin seams of calcareous or sandy shale. At the type locality the estimated thickness of the middle member is about 95 feet.

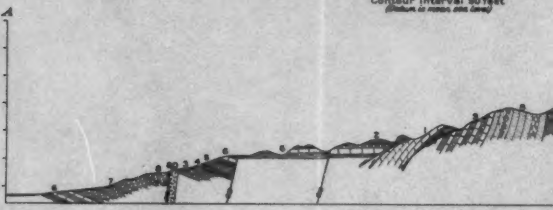
The upper member of the Gabbs formation is about 35 feet thick. Lithologically it is inseparable from the underlying middle member of the Gabbs formation and the overlying Jurassic Sunrise formation. This upper member was differentiated exclusively on the basis of fossils and is thus, properly speaking, a faunal zone rather than a lithologic member.

Fossils and age.—Abundant and well-preserved marine invertebrates were collected from the Gabbs formation in the Gabbs Valley Range.

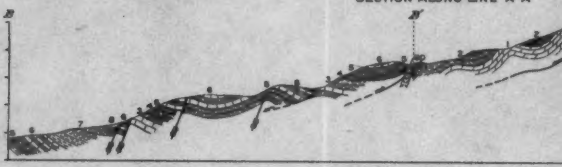




MAP
Contour interval 50 feet

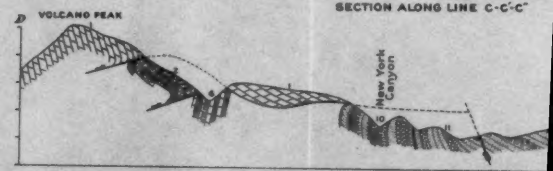


SECTION ALONG LINE A-A'



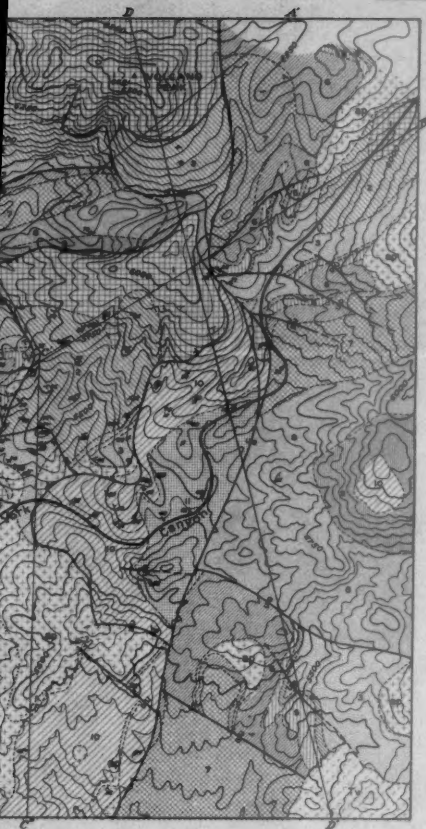
SECTION ALONG LINE B-B'-B"

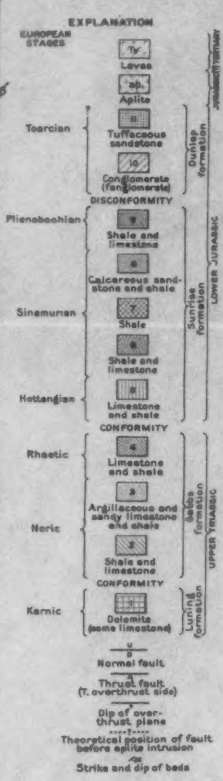
SECTION ALONG LINE C-C'-C"

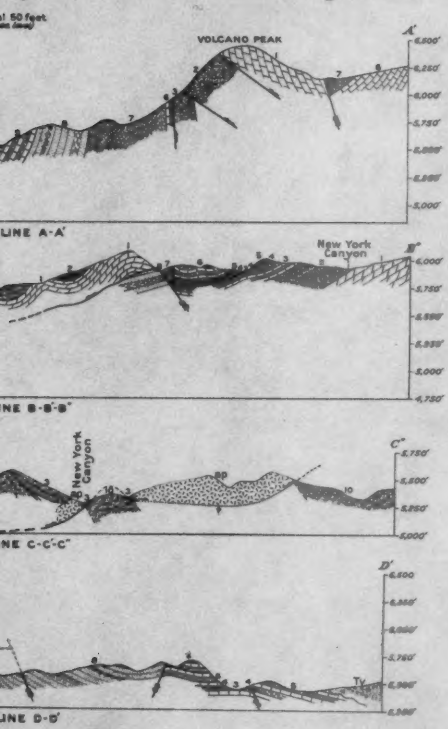


SECTION ALONG LINE D-D'

GEOLOGIC MAP OF NEW YORK CANYON, G







...YON, GABBS VALLEY RANGE.

Gabba formation

TABLE 4.—*Stratigraphic distribution of fossils of the Gabbs formation*

		UPPER TRIASSIC								
		Karnic	?	Noric						Rhaet.
			Discophyllites patens zone	*Sagenites giebeli* zone	*Cladiscites ruber* zone	*Cyrtopleurites bicrenatus* zone	*Pinacoceras metternichi* zone	*Sirenites argonautae* zone	*Choristoceras marshi* *Pteria contorta* zone	
Gabbs formation top	*Pteria* aff. *P. contorta* (Portlock)*						-		x	
	Choristoceras marshi Hauer								x	
	C. cf. *C. ammonitiforme* Gümbel								x	
	Arcestes cf. *A. acutegaleatus* Mojsisovics						x			
	Arcestes gigantogaleatus Mojsisovics						x			
	A. intuslabiatus Mojs. (= *A. nevadanus* Hyatt and Smith)						x			
	A. oligosarcus Mojsisovics						x			
	Stenarcestes peribothrus Mojsisovics						x			
	Cladiscites tornatus Bronn						x			
	Paracladiscites multilobatus (Bronn)						x			
	Pinacoceras metternichi (Hauer)						x			
	Placites symmetricus Mojsisovics						x			
	Choristoceras cf. *C. zlamense* Mojsisovics						x			
	Celtites arduini Mojsisovics						x			
	Rhacophyllites debilis (Hauer)			x	x	x	x			
	Proclydonautilus spirolobus (Dittmar)			x	x		x			
	Gonionautilus securis (Dittmar)						x			
	Sagenites cf. *S. princeps* Mojsisovics			x						
	S. cf. *S. giebeli* (Hauer)			x						
	Rhabdoceras suessi Hauer		x	x	x		x	x		
	Cochloceras fischeri Hauer		x	x	x					
bottom	*Rhacophyllites* aff. *R. debilis* (Hauer)			x	x	x	x			

* = "Avicula" contorta Portlock.

The fossils, although occurring throughout the entire thickness of the formation, are particularly common in three belts: (1) Near the base of the lower member (*Sagenites giebeli* fauna), (2) near the top of the lower member and throughout the middle member (*Pinacoceras metternichi* fauna), and (3) in the lower part of the upper member (*Choristoceras* fauna).

The lowest assemblage contains:

Sagenites cf. *S. princeps* Mojsisovics
S. s. cf. *S. giebeli* (Hauer)
Rhabdoceras suessi Hauer
Cochloceras fischeri Hauer
Rhacophyllites cf. *R. debilis* (Hauer)

This fauna is correlated with the *Sagenites giebeli* zone of the Lower Noric stage of the Upper Triassic of Europe. This zone has not been previously recognized elsewhere in North America. It may correspond in time to the coral zone of the Shasta region, California, and also possibly in part to the *Halorella*-bearing beds in the Santa Ana Mountains, California.

The fauna from the middle of the Gabbs formation also show a striking kinship to forms of the Mediterranean region (Table 4). Most of the forms obtained from this zone are identical with European species, and only a small part of the fauna consists of forms which are not conspecific with any of the known foreign species. This assemblage matches very closely the fauna from the *Pinacoceras metternichi* zone of the upper Noric stage in Europe, notwithstanding the fact that the zonal fossil itself—*Pinacoceras metternichi*—is exceedingly rare in the Nevada assemblage. This fauna as a distinct assemblage was first recognized by Stanton (1926, p. 485), who listed from it such ammonoids as *Cladiscites* aff. *C. tornatus, Rhabdoceras,* and *Rhacophyllites* and correctly assigned it to the Noric.

The third fauna was obtained from the upper member of the Gabbs formation. Although the fossils here are less abundant and not so well preserved as those from the middle of the formation, nevertheless they appear to be sufficiently distinctive and diagnostic to permit correlation with the *Pteria contorta* (*"Avicula" contorta*) zone of the Rhaetic of Europe.

Correlation.—The results of the comparative study of the fossils from the Gabbs formation with the corresponding faunas abroad, together with the apparent stratigraphic continuity within the Gabbs formation, indicate that the Noric, beginning with the zone of *Sagenites giebeli,* and the entire Rhaetic stages of the European section are represented in the formation.

The upper "member" of the Gabbs formation, representing the Rhaetic stage, cannot be correlated with any known Triassic rocks in western North America, as no marine strata of Rhaetic age have been heretofore recognized in this continent. The presence of marine Rhaetic beds in Nevada invalidates the statement that "nowhere on the [North American] continent is there any latest Triassic . . . marine sediment" (Crickmay, 1931, p. 18), and it is now thought likely that further research will result in a discovery of marine Rhaetic strata at other localities in western North America.

In this connection it is of interest to note that Clapp and Shimer (1911, p. 425-428) in a report on the Sutton formation fauna of British Columbia stated that several species from that formation are closely related to the Rhaetic species of Europe. Among these they listed *Choristoceras suttonensis* Clapp and Shimer, *Myophoria suttonensis* Clapp and Shimer, and two species of corals. The remaining several species of corals they recognized, however, to be related to the Jurassic species, and for this reason the entire assemblage was dated as early Jurassic (Liassic) in age. A few years later these same corals and the associated mollusks were referred by Martin (1916, p. 709) to the lower Noric stage of the Upper Triassic. In this conclusion Martin was followed by Smith (1927, p. 10) who also was of the opinion that the Sutton formation corals represented the interregional coral zone of early Noric age. The fact that the very same corals have been regarded by some workers as Jurassic while others thought them to be of Triassic age only serves as an additional proof that the reef-building corals cannot be relied upon in exact dating of the early Mesozoic strata (Muller, 1936).

The genus *Choristoceras* is abundantly represented in the upper part of the Gabbs formation, in the beds which correspond to the Rhaetic and the upper part of the Noric stages. In British Columbia *C. suttonensis* is represented by only a fragment of a whorl which does not permit accurate comparison with the known species of the genus, but even though specific affinities of the species *C. suttonensis* cannot be established it is nevertheless more probable that it is of Rhaetic or late Noric age and not of early Jurassic or early Noric age as has been held by previous authors. In the Gabbs formation fauna there is also a species of *Myophoria* which is very similar although not conspecific with *M. suttonensis*. It occurs near the top of the Gabbs formation but, unfortunately, at this particular locality it could not be determined whether the beds in question are uppermost Noric or Rhaetic.

From the foregoing it appears that there is only suggestive evidence that the Sutton formation of British Columbia might correlate with the upper part of the Gabbs formation, but pending further investigation this

correlation cannot be regarded as established with any degree of certainty.

The middle and the lower members of the Gabbs formation also cannot be correlated on the basis of their faunas with any known Upper Triassic formation in western North America. However, it should be pointed out that in the Alpine region the Noric fauna, very similar to that of the Gabbs formation, and containing such distinctive ammonites as *Arcestes gigantogaleatus* Mojsisovics and *A. acutegaleatus* Mojsisovics (commonly referred to as the "galeati" fauna) has been shown to be a faunal facies of the *Monotis salinaria* (*Schlotheim*) assemblage (Arthaber, 1906, p. 344-345). Still another facies contemporaneous with the "galeati" fauna is represented by the reef-building corals. The American West Coast pelecypod *"Pseudomonotis" subcircularis* (Gabb) is now known to be generically, and in part specifically, identical with the Alpine *Monotis salinaria* (Muller, 1938, p. 1893) and, in all probability, represents a different faunal facies of the "galeati" fauna of the Gabbs formation. The absence of *"Pseudomonotis" subcircularis* in the Hawthorne and Tonopah quadrangles is therefore not an indication of the incompleteness of the Upper Triassic sequence in that area. It merely means that the realm of the *Monotis* or *"Pseudomonotis"* faunal facies did not extend into this area at the time when the "galeati" ammonites formed the dominant faunal element.

The striking analogy of the late Triassic (Noric) faunal facies in the Mediterranean and in western North America may be used as a basis for correlating the *"Pseudomonotis"*-bearing beds with the strata characterized by the so-called "galeati" fauna. The validity of such a correlation is further substantiated by the fact that the *"Pseudomonotis"*-bearing beds in California, Nevada, and elsewhere in western North America overlie the uppermost Karnic *Tropites*-bearing beds and are therefore Noric in age—*i.e.*, of the same age as the lower and middle members of the Gabbs formation.

For these reasons the lower two members of the Gabbs formation are to be correlated, although perhaps only in part, with the Brock shale and Swearinger slate of California, both of which carry abundant remains of *"Pseudomonotis" subcircularis* and overlie the Hosselkus limestone of Karnic age. Other formations which correlate with the Gabbs formation, at least in part, are the *"Pseudomonotis"*-bearing slates in Muttleberry Canyon, West Humboldt Range, and the slates in Eldorado Canyon, near Dayton, Nevada, as well as other *"Pseudomonotis"*-bearing beds in British Columbia, Alaska, and elsewhere.

In the Santa Ana Mountains of Southern California and in central Oregon there are Triassic sediments which contain the brachiopod genus

Halorella. In other parts of the world the genus *Halorella* is known only from the rocks of Noric age, and for this reason the rocks in question may also be regarded as roughly equivalent to a part of the two lower members of the Gabbs formation.

The presence in the Gabbs formation of the cephalopod fauna which heretofore has been regarded as typical or peculiar to the Mediterranean province is of paramount significance. It confirms the view expressed by Spath in 1930 (p. 87) that "ammonite faunas, when they are strictly contemporaneous, are remarkably similar wherever they occur."

It also sheds a new light on the sequence of geologic events in western North America at the close of the Triassic period as they were understood by previous authors. The closing chapter of the Triassic period was described by Smith (1927, p. 3) as follows:

> "In the time represented by the *Pseudomonotis* zone the European connection was cut off, and intermigration took place through the Bering portal, which opened into the Arctic sea. This fauna came down on both sides of the north Pacific, and spread southward even below California, probably under influence of a Boreal current.
>
> "With this epoch the marine Triassic history of North America ends, for everywhere, from Alaska to California, the oldest Jurassic beds lie unconformably upon the upturned edges of the Triassic."

Crickmay (1931, p. 18) in a discussion of the Jurassic history of North America states that

> "it is generally assumed that at the end of the Triassic, the ancestral North American continent of that time stood well above the sea level. This idea is based mainly on the fact that nowhere on the continent is there any latest Triassic or earliest Jurassic marine sediment."

In the light of the new evidence presented in this article there appears to be no indication that the connection between the West Coast of North America and the Mediterranean region was severed near, or at the close of, the Triassic period. On the contrary, there is positive proof, at least for west-central Nevada, that the late Triassic seas of that area persisted, perhaps with minor oscillations, well into the beginning of the Jurassic and that the commingling of the Mediterranean and western American faunas continued without apparent interruption through Karnic, Noric, Rhaetic, and Liassic ages.

JURASSIC SYSTEM

SUNRISE FORMATION (LOWER JURASSIC)

Definition.—The Sunrise formation derives its name from Sunrise Flat, in the Gabbs Valley Range.

The Sunrise formation is best exposed in the vicinity of New York Canyon, in the southwest part of the Gabbs Valley Range, south of Volcano Peak. In this area the Sunrise formation is very intricately folded and faulted (Pl. 4) except for the small area about 2 miles from

the mouth of New York Canyon selected as the type locality of the formation. Here a homoclinal section of the Sunrise strata is readily accessible and presents an unbroken sequence from the base of the formation to the overlying conglomerates of the next younger Dunlap formation.

The Sunrise formation at the type locality cannot be readily separated on the basis of lithology from the underlying Gabbs formation of late Triassic age. Furthermore, in this area there is no apparent physical break between the rocks of the two systems, so that the Triassic-Jurassic boundary had to be drawn arbitrarily in the 30 feet of unfossiliferous shales and limestones which intervene between the uppermost fossiliferous Triassic beds below and the lowermost fossiliferous Jurassic beds above (Pl. 5, fig. 2).

The upper boundary of the Sunrise formation is marked by an abrupt change from fine clastic and calcareous marine sediments to overlying conglomerate (fanglomerate), tuffs, volcanic breccias, and lavas of the Dunlap formation. In spite of this abrupt change in lithology the contact between the Sunrise and the Dunlap formations at the type locality appears to be conformable. Elsewhere within the area of the two quadrangles the basal conglomerate of the Dunlap formation contains boulders derived from the Sunrise and other subjacent units, indicating that an erosion interval separates the two formations throughout most of the area.

Other exposures of the Sunrise formation showing similar lithologic character are found in the southwestern part of the Shoshone Mountains, southwest of Grantsville, in isolated patches in the Garfield Hills, between Garfield Flat and Pamlico, in the Paradise Range, and in a very small area in the northwestern Pilot Mountains.

Lithologic character.—The Sunrise formation consists dominantly of interbedded shales and limestones, with only slight variation in the proportion of each throughout the section. The shales are partly altered to slate and locally to hornfels, whereas the interbedded limestones are unaltered—for the most part argillaceous, cherty, arenaceous, and in places grading into calcareous sandstone. In the lower part of the formation dark-brown oölitic limestone from 10 to 15 feet thick served as a "key bed" in working out the structure. In general the Sunrise formation is lithologically similar to the underlying Gabbs, and in deformation both behaved as an incompetent unit, in sharp contrast to the more massive Luning formation.

The Sunrise formation at the type locality in New York Canyon, Gabbs Valley Range, has been subdivided into five members designated by symbols which also appear on the map (Pl. 4).

FIGURE 1. WEST FRONT OF GABBS VALLEY RANGE
(Photo by Eliot Blackwelder)

FIGURE 2. TRIASSIC-JURASSIC TRANSITION BEDS NEAR VOLCANO PEAK
Lowest fossiliferous Jurassic beds are near middle of slope, where man is sitting. Outcrop of Upper Triassic beds (*Pteria contorta* zone) shown at bottom of photograph. Between is about 30 feet of unfossiliferous shales.

GABBS VALLEY RANGE

From top to bottom, and beginning at the contact with the conformably overlying Dunlap formation, the type section is as follows:

Symbols used on map	*Description*	(Feet)
9	Black shales, argillaceous limestones, with cherty nodules. Fossils rare	250
	Reddish-gray sandy shale with fragments of *"Deroceras"* and *Trigonia*	10
8	Grayish-brown calcareous sandstone and shale locally grading into argillaceous limestone. *Trigonia* cf. *T. clavellata* Sowerby, *Entolium* sp. *Pholadomya nevadana* Gabb, *P. multilineata* Gabb, *Weyla alata* (Buch), and *W.* cf. *W. bodenbenderi* (Behrendsen)	400±
7	Dark-gray slaty shales—light-gray upon weathering. Subordinate limestone, somewhat metamorphosed. Abundant impressions of *"Arietites"*-like ammonites not recognizable generically	400±
	Black to brown sandy shale and cherty limestone. Fragments of crinoids and a *Rhynchonella*-like brachiopod	100±
6	Dark-brown oölitic limestone, massive and coarsely crystalline. Richly fossiliferous. *Ammonites* cf. *A. rotiforme* Sowerby, *A. bisulcatus* Brugiere, *Megarietites meridionalis* (Reynes), *Ammonites rotator* Sowerby, *Tmaegoceras* (several species), *Pecten acutiplicatus* Meek, *Stylophyllopsis* sp., and *Pholadomya nevadana* Gabb	10
5	Argillaceous and cherty limestone interbedded with shale. Limestone is dominant near the top, whereas shale forms greater proportion of the thickness near the base. Richly fossiliferous; *Psiloceras* (several species), *Euphyllites, Waehneroceras, Schlotheimia, Phylloceras,* and *Pecten* aff. *P. textorius* Quenstedt	75
	Total thickness	1245

This section is underlain conformably and with no apparent physical break by the Upper Triassic Gabbs formation.

Fossils and correlation.—The lowest assemblage from the Sunrise formation is characterized by species of *Psiloceras, Waehneroceras,* and *Euphyllites,* with which are associated *Pecten* aff. *P. textorius* Quenstedt and *Goniomya* cf. *G. quenstedti* Neumayr. These indicate the base of the Hettangian stage of the Lias of the European section corresponding to the Psiloceratan "age" of Spath, Buckman, and others, which is the lowest Jurassic division recognized in Europe and other parts of the world.

Considerable importance is attached to this part of the Sunrise section, as marine rocks of earliest Jurassic age have not been heretofore recognized in North America.

The stratigraphically higher assemblage comes from oölitic beds 75 to 85 feet above the base of the formation and contains among other forms:

Megarietites meridionalis (Reynes)
Ammonites bisulcatus Brugiere
A. rotator Sowerby

Tmaegoceras[4] (several species)
Pecten acutiplicatus Meek
Stylophyllopsis sp.
Pholadomya nevadana Gabb

According to the known occurrences of similar and identical species in Europe, this member of the Sunrise formation is correlated, in part, with the *Ammonites bucklandi* zone or the base of the Sinemurian stage of the Lias.

Vermiceras crossmani Hyatt, *Coroniceras claytoni* Hyatt, and *Arnioceras nevadanum* (Gabb) reported by Hyatt (1894, p. 400) from "Volcano" district in all probability came from this part of the section.

There is a strong suggestion that the oölitic bed, containing side by side such forms as *Ammonites bisulcatus, Coroniceras,* and *Arnioceras,* represents a condensed deposit of more than one zone. The analysis of this assemblage will be discussed at greater length in the forthcoming paleontological report.

This part of the Sunrise formation is tentatively correlated with the lower part of the Harbledown formation of British Columbia, with the Modin formation in Shasta County, and with a part of Milton formation (believed to be the equivalent of the Sailor Canyon) in Plumas County, California. It was recently shown by Crickmay (1931, p. 23) that the Sailor Canyon formation, originally assigned to Jura-Trias (Turner, 1894, p. 228-232) and later shifted to Triassic, has yielded a specimen of *Arniotites* sp. and hence must be, at least in part, equivalent to the Sinemurian.

The Trail formation of the Mount Jura region as originally described by Diller (1908, p. 35) but not as redefined by Crickmay (1933, p. 903) may also be provisionally assigned to this part of the section.

Other areas in Nevada where rocks equivalent to the Sinemurian are known to occur are: Muttleberry Canyon, Humboldt Range, and possibly El Dorado Canyon, south of Dayton, Nevada.

The third faunal assemblage, found near the top of the Sunrise formation (divisions 8 and 9), contains:

Eoderoceras ex gr. *E. armatum* (Sowerby)
"*Deroceras*" cf. "*D.*" *densinodum* Quenstedt
Weyla alata (Buch)
Trigonia aff. *T. clavellata* Sowerby
T. sp.
Entolium cf. *E. meeki* Hyatt
Pholadomya nevadana Gabb
P. multilineata Gabb
Rhynchonella sp.

This assemblage marks the lower part of the Pliensbachian (= Charmouthian) stage, corresponding to the "Deroceratan age" of Buckman

[4] The genus *Tmaegoceras* was kindly identified by Dr. L. F. Spath of the British Museum (Natural History).

and others. It should be borne in mind, however, that according to Spath (1925, p. 167-172) the deroceratids, having greater stratigraphic range than was formerly believed, are not satisfactory for accurate dating of strata. It is therefore possible that this uppermost Sunrise fauna may prove to be slightly older—*i.e.*, of late Sinemurian age.

The Lilac formation (Crickmay, 1933, p. 896-897) and the Hardgrave formation (Diller, 1908, p. 39) of Mount Jura, Plumas County, California, are approximately of the same age as the upper part of the Sunrise formation. The large pectens from the Lilac formation described by Crickmay as *Parapecten praecursor* (Crickmay, 1933, p. 905, Pl. 25, figs. 4 and 5 but not fig. 6) and *Parapecten acutiplicatus* (Meek) from the Hardgrave formation (Crickmay, 1933, p. 906, Pl. 25, figs. 1-3) appear to be nothing more than the common Liassic form, *Weyla alata* (Buch), also known as *"Vola" alata*. The specimen figured by Crickmay as Plate 25, figure 6, also appears to be misidentified. It is most likely a left valve of Meek's *Pecten acutiplicatus*, the type of which has been recently examined by the senior writer at the Museum of Comparative Zoology through the courtesy of Dr. Percy E. Raymond. Both *Weyla alata* and *Pecten acutiplicatus* are rather common in the upper part of the Sunrise formation.

Triassic-Jurassic contact.—As has been already mentioned there is no perceptible change in lithology between the uppermost beds of the Gabbs formation of late Triassic age and the lowermost beds of the overlying Sunrise formation of earliest Jurassic age. In the Gabbs Valley Range between the uppermost fossiliferous Triassic beds and the lowermost fossiliferous Jurassic strata there is about 30 feet of dark unfossiliferous shales with occasional thin beds of impure limestone.

For the detailed geologic mapping (Pl. 4) which was carried on in the vicinity of New York Canyon the boundary between the Triassic and Jurassic systems was arbitrarily established at about 20 feet above the uppermost occurrence of Triassic fossils and at about 10 feet below the *Psiloceras*-bearing beds, which represent the base of the Lower Jurassic (Fig. 3; Pl. 5, fig. 2). At this horizon there is a barely perceptible change in the color of the shales from dark, purplish-gray below to slightly darker above.

Similar relationship between the Triassic and Jurassic was also observed in the Garfield Hills, northeast of Pamlico.

Conformity between the marine uppermost Triassic and the lowermost Jurassic has not been generally recognized in North America. In the past it was generally believed (Smith, 1927, p. 3) that

"everywhere, from Alaska to California, the oldest Jurassic beds lie unconformably upon the upturned edges of the Triassic."

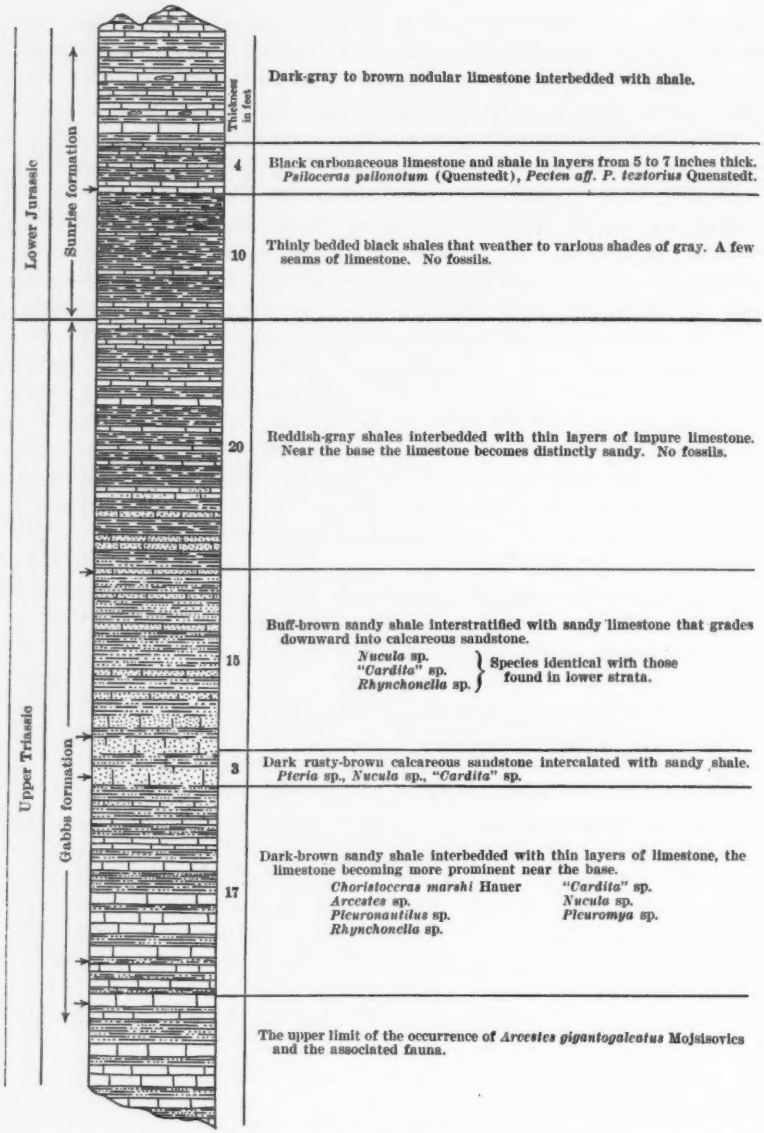

Lower Jurassic
Sunrise formation
Upper Triassic
Gabbs formation
Thickness in feet
Dark-gray to brown nodular limestone interbedded with shale.
4
Black carbonaceous limestone and shale in layers from 5 to 7 inches thick. Psiloceras psilonotum (Quenstedt), Pecten aff. P. textorius Quenstedt.
10
Thinly bedded black shales that weather to various shades of gray. A few seams of limestone. No fossils.
20
Reddish-gray shales interbedded with thin layers of impure limestone. Near the base the limestone becomes distinctly sandy. No fossils.
15
Buff-brown sandy shale interstratified with sandy limestone that grades downward into calcareous sandstone.
Nucula sp.
"Cardita" sp.
Rhynchonella sp.
Species identical with those found in lower strata.
3
Dark rusty-brown calcareous sandstone intercalated with sandy shale. Pteria sp., Nucula sp., "Cardita" sp.
17
Dark-brown sandy shale interbedded with thin layers of limestone, the limestone becoming more prominent near the base.
Choristoceras marshi Hauer
Arcestes sp.
Pleuronautilus sp.
Rhynchonella sp.
"Cardita" sp.
Nucula sp.
Pleuromya sp.
The upper limit of the occurrence of Arcestes gigantogaleatus Mojsisovics and the associated fauna.

FIGURE 3.—*Columnar section showing Triassic-Jurassic transition beds; near Volcano Peak, Gabbs Valley Range*

The unconformable relationship between the Triassic and Jurassic, believed to exist in Muttleberry Canyon, West Humboldt Range, Nevada, was for many years regarded by practically every worker in the Mesozoic as typifying the general condition presumably existing everywhere in the Northwest. A field study of the Muttleberry Canyon section by the senior author failed to reveal any apparent stratigraphic break in a homoclinal section between the uppermost Triassic *"Pseudomonotis"* beds and the *Arnioceras*-bearing beds of the Lower Jurassic. At this locality the *"Pseudomonotis" subcircularis* beds can be followed upward into the fossiliferous beds of early Lower Jurassic age without any noticeable change in lithology or attitude.

Furthermore, it can be seen that Louderback (1904, p. 335) was by no means certain about the unconformity between the Triassic and Jurassic in the West Humboldt Range, when he wrote that

"some deformation and erosion *probably*[5] took place at the end of the Triassic, but of comparatively limited magnitude, for Jurassic is disturbed and folded and altered to about the same extent that the Triassic is."

This statement appears to be the only basis for the concept of an unconformity between the Triassic and Jurassic of the Western States, which has become so firmly established in the literature that it has completely overshadowed the earlier conclusion of Clarence King (1878, p. 269) that in the West Humboldt Range "the Trias and Jura are perfectly conformable." The burden of proof therefore still rests with those who claim that the contrary is the case.

In eastern Oregon, Packard and Lupher have established an erosion interval between the Triassic and Jurassic for that area but, according to Clark (1930, p. 50-51), in the Milton formation (believed to be equivalent to the Sailor Canyon formation) of Plumas County, California,

"no evidence of an unconformity was seen between the Noric and early Lias fossil horizon. As far as could be determined the Jurassic is conformably on the Triassic."

Crickmay (1928, p. 54-58) in his study of the stratigraphy of the Parson Bay region, British Columbia, apparently likewise had some difficulty in separating the Triassic system from the Jurassic. The Triassic argillites of the restricted Parson Bay formation appear to resemble closely the argillites of the Harbledown formation of early Jurassic age. It is also stated that the fossil zones are too widely separated to be of use in finding the exact boundary between the two systems. Crickmay observes, however, that a change in lithologic character takes place about

[5] The writer's italics.

400 feet above the occurrence of the *"Pseudomonotis" subcircularis* Gabb. Crickmay says:

"At this horizon the argillite loses entirely its calcareous character. Then follow about two hundred feet of black argillite with many thin beds of green cherty quartzite which gives the rock a strongly banded appearance."

The presence of the ammonite genus *Melanhippites* in these banded strata indicates that these rocks belong to the Jurassic. Crickmay continues further that

"it would seem that the base of the green banded beds is the local base of the Jurassic system. This contact seems perfectly conformable, but as there is a gap in the record it should be called a disconformity."

This, as Crickmay remarks, is "an inference based upon fossil evidence alone." It seems doubtful, however, if these fossils, in zones several hundred feet apart, can be used at all for establishing erosional gaps in the intervening unfossiliferous but conformable sequence of beds, and it cannot be regarded as proved that

"the latest epoch of the Triassic, the Rhaetic, is not represented by deposits, nor are the earliest ages of the Jurassic—Psiloceratan to Coroniceratan."

On the basis of field evidence presented by Crickmay it also does not seem warranted to assert that

"gentle uplift and some erosion mark latest Triassic and earliest Jurassic time at this locality. The relations [are] very similar to those in Western Nevada."

The erosional gap between the Triassic and Jurassic systems in the Parson Bay region is thus by no means an established fact. From Crickmay's report it appears more likely that the relationship is a conformable if not a gradational one.

Concerning the continental Triassic and Jurassic deposits in the interior of the United States, no one seems to know where the Triassic-Jurassic boundary lies, and it also appears to be impossible to say whether in that area latest Triassic and earliest Jurassic are present or absent (Baker, Dane, and Reeside, 1936, p. 55-58).

The evidence cited in this paper shows that in the Hawthorne and Tonopah quadrangles the marine Triassic strata have not been deformed and bevelled prior to the deposition of early Jurassic sediments. Near the contact the strata of the two systems are perfectly conformable (Pl. 5, fig. 2) without a trace of any physical evidence to suggest that either an interruption in sedimentation or an erosional interval occurred between the Triassic and Jurassic.

DUNLAP FORMATION (LOWER JURASSIC)

Definition, lithologic character, and origin.—The Dunlap formation—the youngest Mesozoic sedimentary unit in the area—was for the most

part deposited after the emergence and the beginning of folding of the older Mesozoic formations.

The lithologic composition varies greatly from place to place. Definite marine deposits, including thin beds of limestone, are sparingly represented. Much of the formation consists of sandstone, commonly coarsely cross-bedded with locally thick conglomerates and fanglomerates whose pebbles are of local derivation. In certain areas the upper part of the formation consists largely of volcanic rocks with interbedded sandstone, in part tuffaceous.

The principal areas of the Dunlap formation lie within a broad belt bordering the southern margin of the Luning formation. In the southernmost areas, as in the Excelsior Mountains, the Dunlap overlies and is surrounded by the Excelsior formation. In the Pilot Mountains and Garfield Hills the Dunlap overlies the contact of Excelsior and Luning. The contact of Dunlap with Excelsior shows a marked angular unconformity; in the Pilot Mountains this is as much as 90 degrees. The contact with the Luning also shows angular discordance but rarely over 20 degrees.

In the northern part of this belt, a few miles from the margin of the Luning area, the Dunlap formation overlies the Sunrise, either with apparent conformity, as in the northwestern Garfield Hills and New York Canyon, or with only moderate erosional unconformity, as in Sunrise Flat.

Outside the major belt in the south, small areas of Dunlap formation in the Paradise Range rest on the Permian (?) and in the Shoshone Range on the Sunrise formation.

In certain areas, such as the Gabbs Valley Range, the southwestern Garfield Hills, and the Excelsior Mountains, the sedimentary part of the Dunlap is overlain by several hundred feet of volcanic rocks, including greenstones, greenstone breccias, felsites, and tuffs with interbedded tuffaceous sandstones. These are petrographically indistinguishable from similar rocks in the Excelsior formation. The volcanic portion of the Dunlap, however, lacks the massive cherts common in the Excelsior and contains a larger proportion of interbedded sandstones.

In addition to the basin or basins which border the margin of the Luning area and are indicated by the present distribution of the formation and its relation to the Sunrise and Luning in different areas, there is a suggestion that in certain parts of the marginal areas small troughs were formed during the initial stages of more intense folding. Such local troughs may have been formed in the Gabbs Valley Range prior to volcanism and in the Pilot Mountains and the northeastern Garfield Hills probably later than the volcanic phase. In these areas the lower

part of the section consists chiefly of sandstone with a few beds of marine limestone and beds and lenses of conglomerate whose pebbles consist entirely of the resistant chert of the Excelsior formation. In the upper part there is a sharp change in lithology, and the formation consists principally of conglomerates and fanglomerates whose material is derived entirely from the Luning formation. As Lawson (1913, p. 332) has pointed out, the presence of fanglomerate is indicative of bold relief and aridity, and bold relief itself is the immediate result of acute diastrophism.

Continued deposition of the Dunlap during the initial stages of thrusting is shown by the presence of conglomerates and fanglomerates derived from the massive Luning dolomite and limestone which form the upper plate of the earlier thrusts in the Pilot Mountains and Gabbs Valley Range. These fanglomerates, similar in all respects to those of the upper part of the Dunlap, are themselves overridden by the Luning limestone and dolomite from which their pebbles are derived, and similar conglomerates and fanglomerates overlie unconformably the rocks of the upper plates of the thrusts.

Typical sections.—The following summary descriptions of typical sections of the Dunlap formation in the Pilot Mountains, Gabbs Valley Range, and the southeastern part of the Garfield Hills will give some idea of the varied lithology and changing conditions of deposition of the formation.

The type section of the Dunlap formation is in the western part of the Pilot Mountains in the vicinity of Dunlap Canyon. Here as much as 5000 feet may be present, with the upper part concealed by overthrusting of the Luning formation on the north.

The lower contact with the previously folded chert of the Excelsior formation is very irregular. The angular discordance is in places as much as 90 degrees, and there is nearly everywhere a basal conglomerate of angular chert pebbles. In places the basal conglomerate is lacking and in general it is less than 100 feet in thickness; in a few places dominantly conglomeratic material may persist to 300 feet or more above the base. In the upper part the pebbles are smaller and more rounded, and the conglomerate contains lenses of red sandstone.

The lower half of the formation consists principally of red cross-bedded sandstone with beds of red sandy shale and brown quartzite. Conglomerates mostly with small rounded chert pebbles occur here and there, and in the upper part of the sandstone member conglomerates with pebbles of Luning limestone and dolomite become increasingly abundant. Two thin beds of greenstone breccia and a single thin greenstone flow are

also present in the upper part of the sandstone and are perhaps contemporaneous with the volcanic portions of the formation in other ranges.

Near the top of the sandstone there are thin beds of limestone and dolomite, with interbedded slate and sandstone. Individual beds are not traceable across the area, but the carbonate rocks appear to represent approximately the same horizon. If so, the lower sandstone member varies greatly in thickness; near the head of Cinnabar Canyon less than 500 feet of sandstone and conglomerate is present between the Excelsior and the limestone, while 3 miles to the west on the ridge west of Dunlap Canyon the sandstone must have a thickness of about 3000 feet.

The upper member of the Dunlap formation in the Pilot Mountains consists principally of conglomerates, whose pebbles are limestone and dolomite derived from the upper member of the Luning formation, which is here overthrust on the Dunlap.

Conglomerate is also directly associated with the two earliest thrusts in the northwestern Pilot Mountains. The earlier carries a conglomerate of dolomite pebbles between the overriding dolomite of the Luning formation and the shaly limestone of the Sunrise which forms the lower plate. The pebbles though closely pressed together are not noticeably stretched. In places lenses of sandstone dip about parallel to the thrust plate and at a small angle to the overlying Luning from which the conglomerate is derived. Similar "thrust conglomerate" exposed in a small window of the same thrust overlies unconformably the Dunlap formation consisting of red sandstone; wedges of the conglomerate of dolomite pebbles derived from the thrust plate penetrate the underlying sandstone of the Dunlap (fig. 4).

The second and higher thrust carries the more shaly lower limestone of the Luning above the massive dolomite. Here, since the thin-bedded lower member of the Luning yielded only finer clastic material instead of the coarse conglomerate furnished by the more massive dolomite, the material deposited along the thrust surface consists of sandstone with minor conglomerate containing limestone pebbles. The intricate small-scale contortion in contrast to the broader folding of the Luning suggests crumpling before consolidation.

The section of the Dunlap formation in the Garfield Hills south of Garfield Peak also illustrates the type of Dunlap deposition in a locally deepened trough-like area during the folding of the Luning formation. Here the major part of the Dunlap formation occupies a synclinal trough, with Excelsior on the south and Luning on the north. The Dunlap formation is also present in isolated synclinal areas truncating folds in the Luning. The maximum thickness may be as much as 3000 feet.

The lithologic character is highly varied, and individual beds cannot be

matched on opposite sides of the syncline. Bordering the Excelsior contact there is a bed of impure gray limestone less than 100 feet thick. Similar limestone was also found along the southern part of the Luning contact. Above the limestone bordering the Excelsior, the Dunlap on the southern side of the basin is largely red sandstone, in part cross-

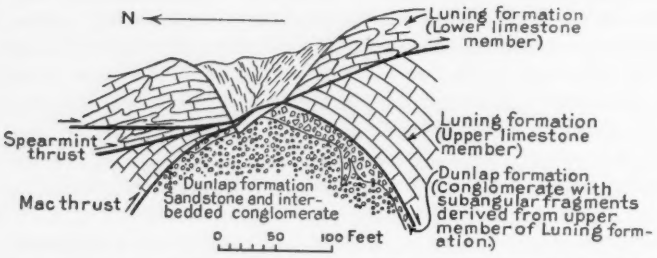

N
Luning formation (Lower limestone member)
Luning formation (Upper limestone member)
Spearmint thrust
Mac thrust
Dunlap formation Sandstone and interbedded conglomerate
Dunlap formation (Conglomerate with subangular fragments derived from upper member of Luning formation)
0 50 100 Feet

FIGURE 4.—*Conglomerate of Dunlap formation below thrust, with wedges penetrating bedding of sandstone (Dunlap) of lower plate. Pilot Mountains*

bedded. In the western part this is varied by a few thin beds of limestone conglomerate, but the great bulk of the material is such as could have been entirely derived from the Excelsior.

Westward from the second ridge and several hundred feet above the base is an andesitic lava flow.

On the northern side of the trough some sandstone is also present, but the Dunlap consists largely of limestone fanglomerate and conglomerate which has a maximum thickness of at least 1000 feet. At the contact with the Luning is a fanglomerate with large angular limestone blocks derived from the Luning formation grading into partly rounded and roughly bedded pebbles a short distance above. The outliers of Dunlap resting on the Luning also consist principally of limestone fanglomerate and conglomerate. In places there is interbedded slate containing scattered limestone pebbles and lenses of limestone conglomerate.

The Dunlap rests on the Sunrise formation with apparent conformity in the New York Canyon area and with erosional, but apparently not angular, unconformity on the Sunrise formation a mile to the east in the ridge west of Sunrise Flat.

In the New York Canyon section the lowest part consists of red cross-bedded sandstone, conformable on shale at the top of the Sunrise formation. A short distance above the base conglomerates with limestone and dolomite pebbles begin to come in and form the major part of the lower 1000 feet of the section. These pebbles are in part derived from the limestone beds of the Sunrise formation but in part from the more massive

limestone and dolomite of the Luning. The upper part of the formation in the New York Canyon area consists dominantly of volcanic material, chiefly tuffaceous sandstone with greenstone breccias in the upper part.

A mile to the east, in the hills southwest of Sunrise Flat, the Dunlap rests on the eroded limestone of the Sunrise formation, indicating about 400 feet of erosion of the Sunrise, but there is no obvious angular unconformity. Here the lower member consists principally of conglomerate with limestone pebbles which interfingers with and is overlain by red sandstone. The upper part of the formation is dominantly volcanic, principally greenstone tuffs, breccias, and flows with a considerable proportion of interbedded sandstones. To the south and east of Sunrise Flat this upper, dominantly volcanic member of the Dunlap is overridden by the dolomite of the Luning formation. Directly below the main thrust and also below subordinate associated thrust planes are lenses of closely compressed dolomitic gravel derived from the Luning of the upper plate. The Luning forming the upper plate is highly folded in places, and overlying it and interbedded with the conglomerate of the Dunlap formation are small lenses of sandstone which show only gentle folding.

Fossils.—A few of the small beds of limestone and dolomite yielded the only identifiable fossils found in the formation.

The *"Arietites"*-like ammonites found in the Excelsior Mountains west of Huntoon Valley are rather poorly preserved but are unmistakably of Lower Jurassic age. In the Garfield Hills many aberrant pelecypods were obtained from a limestone bed near the base of the Dunlap formation. These abnormal bivalves resemble closely the middle Lower Jurassic *Plicatostylus gregarius* Lupher and Packard from central Oregon. The Dunlap formation of the Hawthorne and Tonopah quadrangles is therefore, at least in part, of medial Early Jurassic age. An ammonite, *Harpoceras* sp., was found several hundred feet above the *Plicatostylus gregarius* reefs, indicating that the deposition of the Dunlap formation continued from the Early Jurassic (Pliensbachian) into the early part of late Early Jurassic time (Toarcian). The close proximity of the two formations in the time scale is consistent with the fact that in the New York Canyon area (Gabbs Valley Range) there is locally no apparent stratigraphic break between the Dunlap formation and the conformably underlying Sunrise formation. The highest faunal horizon in the Sunrise is regarded as of early Pliensbachian age, which would indicate that the immediately succeeding strata of the Dunlap formation are either late Pliensbachian or possibly Toarcian.

Other localities at which limestone lenses contain marine fossils are the Pilot Mountains and southern Garfield Hills. Unfortunately, the

type of the material and the nature of preservation does not permit even approximate determination.

Correlation.—The Lower Jurassic rocks of western North America are as yet too little known to permit accurate correlation of any of the units with the Dunlap formation. The lithology and stratigraphic relations of the Dunlap formation indicate that it was deposited during a period of acute diastrophism which began near the middle of early Jurassic time. Lower Jurassic formations of similar lithology and tectonic significance have not been recognized in other parts of this country. It may be stated, however, that rocks of approximately the same age as the Dunlap formation are known in Alaska, British Columbia (Maude formation), in California in Shasta County (Potem formation), and probably in Plumas County (Sailor Canyon formation and Fant formation).

It appears even less feasible to correlate the Dunlap formation with the continental deposits of southern Utah, Arizona, New Mexico, and Colorado. The Lower Jurassic may be represented in this area by the Wingate and Kayenta formations (Baker, Dane, and Reeside, 1936, p. 54-58), but unfortunately their exact stratigraphic position has not yet been determined.

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